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TITLE OF THESIS: The Glaciation of Southwestern Newfoundland

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ABSTRACT:

Southwestern Newfoundland was glaciated by an island-centred ice cap. Labradorean ice did not affect the area. At the Wisconsin maximum, ice flowed into the Gulf of St. Lawrence. During the eastward retreat of the ice margin, flow directions became more variable. The present shoreline was overlapped by the sea between 13,700 and 12,600 years B.P. A tentative age of 13,000 years B.P. is assigned to local readvances of the ice margin into the sea around St. George's Bay. In fjords and along open coasts further north, a stand of the ice margin occurred at 12,600 years B.P. Postglacial marine features rise along the northeast trend of the shoreline. Isobases on an hypothetically synchronous marine limit plane show that isostatic tilt was up towards the northeast. The strong influence of the Newfoundland ice cap is suggested to have been responsible for this. Models are derived which show the nature of postglacial land and sea-level changes in southwestern Newfoundland.

THE GLACIATION OF SOUTHWESTERN NEWFOUNDLAND

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A dissertation submitted to the Faculty of Graduate Studies and Research in partial fulfillment of the requirements for the degree of Doctor of Philosophy.

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PREFACE

This dissertation reports on an investigation of the glacial landforms and deposits of southwestern Newfoundland that was made between 1964 and 1968. Those conclusions which rank as original contributions are, as follows:

First, conclusive evidence was found that the area was not glaciated from Labrador by Laurentide ice. The island of Newfoundland supported its own ice cap at the Wisconsin maximum, an idea which was widely held before 1940. In that year, Flint saw fit to postulate that Labrador ice had invaded at least western Newfoundland, and this has since become strongly entrenched in the literature.

Second, radiocarbon dates on marine pelecypod and barnacle shells from southwestern Newfoundland indicate that late-glacial marine overlap of the present shoreline began about 13,500 years B.P. in St. George's Bay, but was delayed until about 12,500 years B.P. in fjords and along steeply-shelving open shorelines farther north. The earliest date for this event in the area is the earliest so far reported from the Canadian Atlantic Provinces.

Third, a reactivation of the ice margin around St. George's Bay, and an ice-front stand at the heads of fjords and along the northern edge of the Lewis Hills were not regionally significant events: local topographic and glaciological conditions were responsible.

Fourth, in the absence of dated and, in most cases, dateable, regressional marine horizons, a model of land- and sea-level changes is derived from the empirical formula of Andrews (1968a). This enables a date of 13,000 years B.P. to be tentatively assigned to the St. George's Bay readvance, which post-dated the onset of marine overlap.

Lastly, the uplift curve derived for the model allows the variable ages of marine limit features to be taken into account in the reconstruction of a regional trend of isostatic warping. This is tentatively shown to slope up towards the northeast at two feet per mile, and it is suggested that the Newfoundland ice cap had a stronger influence on warping than Labrador-derived ice, which probably lay offshore in the Laurentian Channel.

ACKNOWLEDGEMENTS

I am grateful to many persons and institutions for assistance, encouragement, and inspiration in all stages of the work.

John T. Parry, Geography Department, McGill University, supervised the preparation of the dissertation. His rigorous and diligent scrutiny of drafts has been of great value to me in bringing a semblance of order to the chaos. That I found so much of value in his assistance at long range is one measure of the clarity with which he gave it.

I have been most influenced in this study by V. K. Prest, Geological Survey of Canada. The contact with him came late in the day, but early enough for me to spend five days of the 1968 field season with him in Newfoundland. Those days were the making of this study, and were of more value than all the weeks I spent in the previous three summers along the deserted beaches of St. George's Bay.

Several others have been helpful in discussing certain aspects with me. Doug Grant, Geological Survey of Canada, made clear to me the changing relations of land and sea level through his work in the Maritimes. Jim Shearer, Dalhousie University, opened up the sea-bottom to me while sitting on a far-travelled erratic near Stephenville. John Andrews, University of Colorado, helped me to apply his work on uplift curves to my study area.

J. Brian Bird, Geography Department, McGill University, was kind enough to provide logistical and financial assistance for the 1964 field season.

Vincent Conde, Redpath Museum, McGill University, kindly identified some of the shells collected from the marine deposits.

Jacques d'Avignon, Weather Engineering Company Ltd., made it possible for me to view the study area from the air in 1966.

The United Kingdom Department of Scientific and Industrial Research (now Science Research Council) made funds available to me for field work in 1964, as part of a NATO studentship which enabled me to attend McGill University for three years.

The Canadian National Advisory Committee on Geographical Research awarded me a grant of $1,200 to support field work in 1966. The Division of Quaternary Research and Geomorphology, Geological Survey of Canada, provided generous assistance in 1968, under Project No. 680094. I am very grateful to both these agencies.

Further, the Radiocarbon Laboratory of the Geological Survey of Canada ran five dates for me. Without this assistance I would either be deep in financial debt or still struggling to make sense of the stratigraphic evidence.

Wayne Wells, a Newfoundlander from Stephenville, was a good-natured and long-suffering rod-man for two months in 1968. I owe him more thanks for his companionship, I think, than for his able "rodding."

I am very grateful to those patient and conscientious people who were responsible for the mechanics of the thesis preparation. Mrs. Frost, Mrs. Lonsdale, and Mrs. Reichert, at York University, typed the manuscript in its various stages. Hania Guzewska, Gerry Fradsham, and Jane Fraser, of the Cartographic Office, Department of Geography, York University, meticulously worked their miracles upon my rough compilations of maps and diagrams. Paul Carter and his assistants, of the Department of

Instructional Aids Resources, York University, did the photographic work.

The conclusions reached in this study represent a refinement of those reached by many students of glaciation in Newfoundland since the late-nineteenth century. I would like to dedicate this work to all those men, but especially to Richard Foster Flint, upon his retirement from Yale University, and to the memory of Paul MacClintock, late of Princeton University, who died shortly before my thesis was submitted.

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PART ONE

CHAPTER ONE

Introduction

1. THE OBJECTIVES OF THE STUDY AND THE ORGANIZATION OF THE PRESENT WORK

The present work is an attempt to interpret the landforms and surficial deposits whose genesis was related to a sequence of events that occurred during the glaciation of a part of western Newfoundland. By comparison with areas such as the Great Lakes Basin, the St. Lawrence lowland and its borders, and Labrador-Ungava, the island of Newfoundland has received relatively little recent attention from glacialists whose milieu has been northeastern North America. Quite apart from the lack of published regional summaries (but see Prest, 1970, in press), there had never even been a meeting of interested people on the island, or a field excursion to view glacial phenomena, until 1969. There are obvious geographical reasons for this which appear valid until they are applied to areas such as Labrador-Ungava, or even the high Arctic.

The situation is even more surprising when it is learned that many of the early prominent glacialists at some time either examined or wrote about the glacial features of Newfoundland: DeGeer, Fairchild, Tyrell, Daly, Coleman, Flint, and MacClintock each made a contribution.

Whatever the reasons, the regional aspect of glaciation in Newfoundland and the mainland Maritime Provinces was about as well known in the later part of the 1960's as was that of southern Ontario before the publication nearly twenty years before of Chapman and Putnam's seventeen-year study. During the 1960's the investigation of glacial features in southeastern Canada intensified and produced several locally important results. At the present time, these are being drawn together to produce a regional summary that will stand beside those from other regions in terms of its scientific validity (Prest, 1970, in press).

Within the Canadian Atlantic region, the island of Newfoundland occupies an important position for studies of glaciation. As part of a bold easterly salient of the North American continental margin, situated southeast of a major centre of ice outflow, it is there that evidence can be found of the southeasterly extent of the Laurentide ice sheet at the glacial maximum. Also, since the island has summits over 2,000 feet high along its western margin, and a large part of its area stands above 1,000 feet, it might be expected that it supported an ice cap which was glaciologically independent of the Laurentide ice cap at some stage during the Pleistocene. Further, since the island occupies a maritime position, close to the southern limit of glaciation, it is along its shores that evidence of the earliest stages of postglacial marine overlap can be sought. Lastly, because of this distal situation with respect to an ice sheet, Newfoundland should bear evidence of the outer limits of glacio-isostatic depression of the crust.

All of these topics have been investigated in Newfoundland in a somewhat piecemeal fashion since the late nineteenth century but, to date, no comprehensive island-wide picture has emerged. This study, therefore, may be viewed as an extension across the wide Gulf of St. Lawrence of recent work on the mainland of Atlantic Canada. In the latter area, the relative importance of Laurentide and local ice covers has recently received detailed attention (Prest and Grant, 1969). Those authors have also drawn together the evidence of the temporal and spatial progression of late-glacial marine overlap in the Gulf of St. Lawrence and Bay of Fundy. These two topics are central to the present study in southwestern Newfoundland.

The approach that is adopted herein may be called "traditional," in the sense that an historical reconstruction of glacial events is attempted. The content may be called "historical geology," since uniformitarian principles are applied to the reconstruction of past events. But the end in view is "geographical" in the broad sense of the word, since the present surface of the land is that which I am seeking to explain. Many kinds of evidence are brought to bear on the reconstruction; from the disciplines of solid-earth physics, glaciology, sedimentology, geomorphology, climatology, hydrology, palaeontology, and geochemistry. While a geographer is no better qualified to make an integrative interpretation of the disparate threads of evidence than workers in other fields under the Quaternary "umbrella," it is of the essence of his viewpoint that he should be so concerned.

The dissertation has been organized into two parts. The first deals with the regional geologic and physiographic framework; an overview of previous investigations of glaciation in Newfoundland; and a treatment of those aspects of glacial erosion which are of significance to this study. The second part, comprising the main body of the work, is a treatment of the events with which the origins of the surficial deposits and landforms were associated.

2. THE GEOGRAPHICAL SETTING OF SOUTHWESTERN NEWFOUNDLAND

The location of the study area on the island of Newfoundland is shown in figure one. In all, the study area, shown in more detail in Plate I, extends over roughly 3,500 square miles. Investigations were concentrated around the shores of St. George's and Port au Port Bays and as far inland as routes of access and terrain conditions allowed. The study does not include that part of southwestern Newfoundland which lies south of St. George's Bay. That area, particularly the lowland occupied by the Godroy river valley, is of great interest, but the interest aroused by my studies further north came too late for me to extend the field area to the south.

As the investigations in St. George's and Port au Port Bays progressed, the importance of the phase of marine overlap came to be recognized, so the study area was extended to the north. The coastline between Port au Port Bay and the Bay of Islands, the latter bay itself, and the Bonne Bay region were included for the purpose of gathering evidence of marine overlap over a wider area.

The high plateaus and their steep margins were the subject of an earlier study (Brookes, 1964), in which an interest in the effects of glaciation in Newfoundland was first aroused. Evidence of relatively late glacial features, and the significance to this study of the nature of the bedrock which underlies the highlands, inevitably drew me to the latter from the lowland areas.

Southwestern Newfoundland possesses a greater variety of glacial features than any other part of the island. Elsewhere, the glacierization

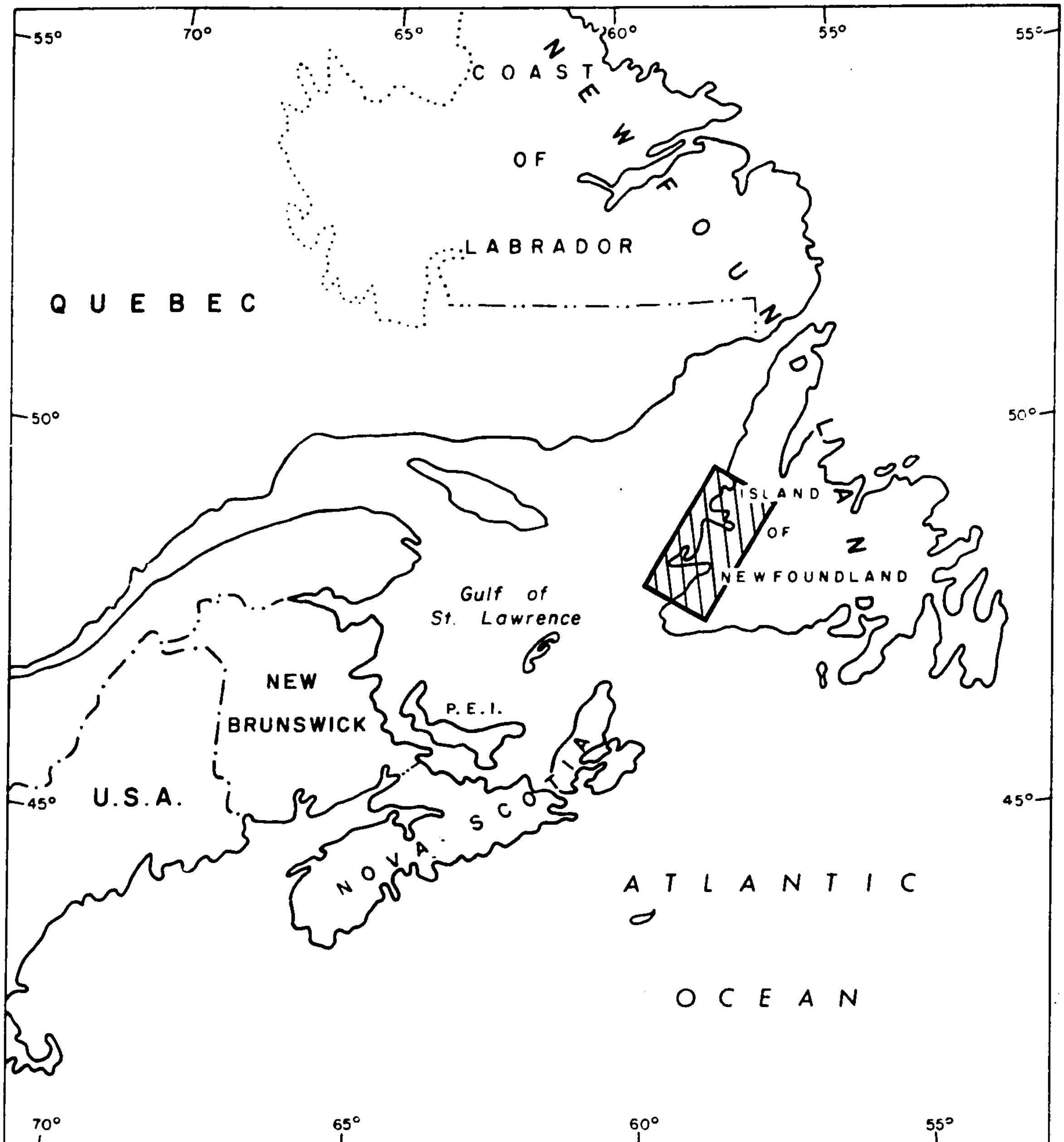
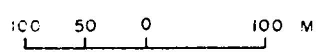


Figure 1.

The position of the island of Newfoundland and the location of the study area.



of plateaus of low to moderate relief has provided little to inspire the glacialist. Even other coasts which were affected by late-glacial marine overlap rarely bear evidence of other events. In southwestern Newfoundland, however, the St. George's Bay lowland is covered by a sheet of glacial deposits which generally thickens seawards from the steep edges of high plateaus. These deposits are excellently exposed to view in coastal cliffs which are being actively eroded, and they are varied enough to indicate several glacial events. Shorelines cut in bedrock are widespread around the Port au Port Peninsula, and along the open shores, and in the fjord bays further north. These littorals are lightly mantled with glacial deposits found in St. George's Bay, so that intra-regional correlations can be made.

The study area is also well endowed with access routes. Most of it is within easy reach of a paved or gravel road. Only in one area, between Port au Port Bay and the Bay of Islands, is it necessary to resort to water transportation and this is easily available at low cost from local fishermen. Settlements of varied size are never out of sight along the coastline. Inland, however, there are practically none. Along the Trans-Canada Highway through the area, there is not a single permanent dwelling between the point where it enters the St. George's Bay lowland and Corner Brook, a distance of nearly 100 miles. Inland of Corner Brook, small settlements occur along the highway in the Humber valley, close to the only centres of employment in Corner Brook, and in Deer Lake, 30 miles away.

Terrain and climatic conditions are not difficult in those parts of the area where the phenomena of interest occur. There is good walking along the beaches and through cleared land close to the coastline. Many shallow, boulder-strewn river channels and dry-weather logging roads offer easy access to deposits exposed further inland.

Summer weather on the west coast of the island is very different from that on the east coast. The difference is due to the warmth of the relatively shallow Gulf of St. Lawrence coastal waters compared to those which are part of the cold Labrador current which maintains small icebergs in east coast coves as far south as St. John's until late June. Some snow patches remain throughout the summer in sheltered sites in the western mountains, but this is more because of heavy winter snowfall than low summer temperatures. Many days in June, July, and August are sunny, with temperatures of 60-70^{o}F, and Corner Brook has a July mean monthly maximum of 88^{o}F (Hare, 1952). Rains are more common in the latter part of August and are associated with the intense cyclonic disturbances which develop from tropical hurricanes that originate in the southeastern United States.

3. BEDROCK GEOLOGY AND PHYSIOGRAPHY

Southwestern Newfoundland is conveniently divisible into three major provinces on the basis of tectonic criteria. Within these provinces, subdivisions can be made according to rock type and physiography. It is the purpose of this section to provide brief illustrated descriptions of these regions as a background to the study of glacial features. Plate II is a simplified geological map of the area, and Plate III shows physiographic subdivisions. Both these plates should be to hand in reading this section.

a. The Grenville Province

The highlands of the Great Northern Peninsula of Newfoundland are built of acid metamorphic rocks, with major intrusions of acid igneous bodies and minor ones of basic composition. These rocks form a roughly rectangular "inlier" which has been assigned to a Grenvillian age (Clifford and Baird, 1962) on the basis of radiometric age determinations which cluster around 1,000 million years B.P.

The Long Range Mountains of the peninsula are bounded on the west by a fault-line scarp that rises from the back of a coastal plain, underlain by steeply inclined lower Palaeozoic sediments, at about 500 feet, to an upland at 2,000-2,500 feet. The steep scarp is notched by steep-sided canyons that are glacially overdeepened preglacial stream valleys. Lakes, such as Western Brook Pond (figure two) and St. Paul's Inlet, are near sea level, and the canyons could be called true fjords if sea level was about 50 feet higher than present.

From the top of the escarpment, at 2,000 feet, the land surface rises gently inland to a line of broad, level summits at 2,500 feet in a series of relatively narrow planation surfaces. Gros Morne, whose summit at 2,651 feet is the second highest on the island, is typical of these highest surfaces (figure three). The surface of the Long Range Mountains slopes gradually eastwards in a series of wide planation surfaces to the shores of White Bay. There, cliffs up to 1,000 feet high border the Atlantic Ocean along another fault-line scarp, parallel to that which bounds the Grenville rocks to the west of the block.

Before glaciation, major rivers drained eastwards and westwards from a divide marked by low saddles between the 2,500-foot summits close

Figure 2: Western Brook Pond: a fjord valley cut into the glaciated upland of the Long Range Mountains of the Great Northern Peninsula of Newfoundland. Cliffs are 2,500 feet high.

Figure 3: The fjord of Bonne Bay from the Trout River road, viewed towards the north. Level skyline summit is Gros Morne, 2,651 feet, second highest in Newfoundland.

Figure 2: Western Brook Pond: a fjord valley cut into the glaciated upland of the Long Range Mountains of the Great Northern Peninsula of Newfoundland. Cliffs are 2,500 feet high.

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to the western front of the block. On the gentler eastern slope, river basins such as that of the upper Humber and the Main River were dendritically organized. Glaciation has had the effect of deepening the western valleys cutting the scarp and of shifting the preglacial drainage divide to the east of those saddles. On the eastern slope, the land surface has been glacially eroded to form a rock-knob topography, with the accentuation of structural weaknesses, particularly faults, in the metamorphic rocks, so that the terrain has taken on the typical glaciated shield aspect. Drainage systems are still organized in well-defined basins, but the pattern has become more angulate.

b. The Taconic Province

The Taconic Province is the most extensive and most varied of the geologic-physiographic regions in western Newfoundland. It has many of the characteristics of a complex orogenic zone and each region within it will be dealt with in the order of its dominance in the landscape.

i. The Long Range Igneous and Metamorphic Complex. The Long Range Mountains of western Newfoundland, excluding those of the Great Northern Peninsula, are built of a complex of igneous and high-grade metamorphic rocks, mostly of acid composition. These represent a zone of intense orogenesis during the mid-Ordovician period (Riley, 1962). Gneisses and granite-gneisses, with large granitic intrusions, comprise the bulk of the rocks outcropping.

The western front of the mountains is marked by a fault-line scarp developed on a Taconic lineament that was strongly reactivated in the Appalachian period of earth movements of Carbo-Permian age. The scarp is not so high as that of the Great Northern Peninsula, although it is developed on a fault that belongs to the same system, but its physiographic expression is no less impressive (figure four). It, too, has been gouged by glaciated canyons, particularly on the east side of the Codroy Valley, outside the area of this study. East of St. George's Bay, the mountain front stands up to 15 miles from the shoreline and is cut by the valleys of major rivers which drain the plateau surfaces to the east into the Gulf of St. Lawrence. These rivers have dissected the scarp and their valleys have been glacially modified (figure five), so that the escarpment appears only as a series of spurs projecting into the sedimentary lowland to the west.

The surface of the Long Range Mountains here is similar to that of those mountains in the northern part of the island. Although the rocks are half a billion years younger, the lithology is similar and both pre-glacial and glacial land-forming processes have been the same. Preglacial planation surfaces slope gently to the east, west and south from a line of 2,000-foot summits standing on a subdued plateau at 1,600-1,800 feet. Glaciation has etched the surface into the typical shield-like terrain of rock knobs and intervening boggy areas with myriad lakes. Some large-scale drainage modifications were effected during the glacial period, and these will be discussed in Chapter Three.

ii. <u>The Bay of Islands Mountains</u>. The Bay of Islands Mountains are four roughly circular massifs of basic and ultrabasic intrusive igneous rocks with minor outcrops of basic volcanic rocks. The four massifs are, from south to north, the Lewis Hills, the Blow me Down (Blomidon) Mountains,

Figure 4: The fault-line scarp of the Long Range Mountains bounding the Codroy Valley on the east, viewed towards the northeast from near Wreckhouse.

Figure 5: The glacial trough of Flat Bay Brook, east of St. George's, viewed towards the southeast from the 950-foot summit of Steel Mountain.

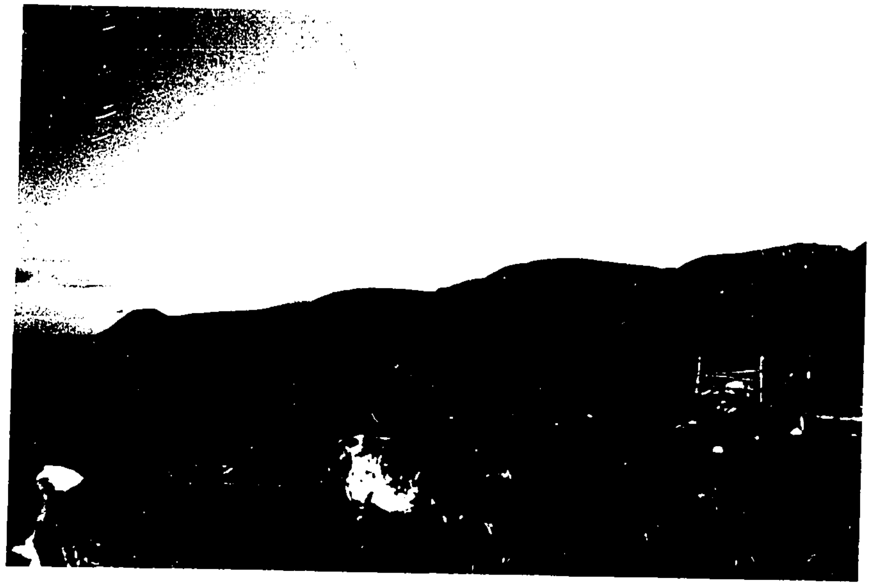

Figure 4: The fault-line scarp of the Long Range Mountains bounding the Codroy Valley on the east, viewed towards the northeast from near Wreckhouse.

Figure 5: The glacial trough of Flat Bay Brook, east of St. George's, viewed towards the southeast from the 950-foot summit of Steel Mountain.

the North Arm Mountain - Mt. St. Gregory block, and Table Mountain, Bonne Bay. They are situated close to the western shore of the island, and in many places steep walls rise out of the sea to the edges of gently rolling summit plains at 2,000 feet. The latter rise more gently to the central parts of the massifs at near 2,500 feet in a series of stepped planation surfaces which radiate out from the highest summits. The steep, often fault-bound, margins are deeply gouged by glaciated canyons (figures six and seven), and many glacial cirques notch the upper slopes.

The rocks which build these masifs are of great significance to this study. In particular, the olivine-rich dunite that is common to each is a valuable indicator of ice-flow directions, since boulders of it are found in the earliest glacial deposits in the Port au Port area.

iii. <u>The Lower Palaeozoic Sedimentary Terrane</u>. Between the western front of the Long Range Mountains, north of St. George's Bay, and the Bay of Islands Mountains, lowlands and low dissected plateaus are developed on Ordovician carbonate and clastic sedimentary rocks. The grey Table Head limestone represents cratonic deposition outside the main Taconic orogenic zone. The clastic St. George Group have been interpreted as belonging to a vast "klippe" which, with the igneous bodies of the Bay of Islands Mountains, was thrust northwestwards from the central part of the island during the Taconic orogeny (Rodgers & Neale, 1963).

Close to the Long Range Mountain front, between Stephenville and Corner Brook, Taconic compression produced thrust slices which are visible today in minor west-facing cuestas. Elsewhere in the sedimentary terrane, structures are limited to northeast-to-southwest trending normal faults

Figure 6: Winterhouse Brook canyon: a glacial valley, two miles long, cut in dunite rock on the northern edge of Table Mountain, Bonne Bay.

Figure 7: The fault-guided, glacial trough of Trout River Ponds, viewed towards the northwest. Nearest lake is nine miles long.

Figure 6: Winterhouse Brook canyon: a glacial valley, two miles long, cut in dunite rock on the northern edge of Table Mountain, Bonne Bay.

Figure 7: The fault-guided, glacial trough of Trout River Ponds, viewed towards the northwest. Nearest lake is nine miles long.

and low-amplitude folds. These sedimentary rocks are weaker than the igneous and metamorphic rocks which border them and many major stream courses are developed in lowlands between higher plateaus. The fjord valleys of Bonne Bay (figure three) and the Bay of Islands, and the valley of Harrys River are major examples of these.

The land surface of the sedimentary terrane is a dissected plateau with low relief near the present shoreline. In the upper reaches of river basins, such as that of Harrys River, deeper dissection has given rise to a local relief of 500-700 feet, with small areas of undissected plateau standing at 1,100 to 1,500 feet above sea level.

c. <u>The Appalachian Province</u>

In southwestern Newfoundland, Carboniferous rocks form the bedrock between the Long Range Mountain front and the shore of the Gulf of St. Lawrence, from the head of St. George's Bay southwestwards to Cape Anguille on Cabot Strait.

Mississippian sandstones, arkoses, and conglomerates of the Anguille Group (Hayes and Johnson, 1938) have been folded into a complex northeast-trending anticlinorium in the Cape Anguille Mountains. The edge of these rocks, in contact with weaker sediments to the north and east, and possibly just offshore, is marked by a steep slope that rises from between sea level and 600 feet on to the relatively smooth surface of the mountains at 1,400-1,750 feet. This smooth surface is actually a series of planation surfaces that step up from the western and eastern margins of the plateau to low residual summits that lie closer to the eastern margin. From level upland valley heads, streams become progressively more entrenched into the plateau margins until they cascade into the sea on the western side, or into the Codroy lowland on the eastern side, in deep, narrow defiles (figure eight).

The Cape Anguille Mountains push a bold nose into a lowland east of St. George's Bay that is underlain by weaker Carboniferous sediments of the Codroy and Barachois Groups. These rocks are also predominantly reddish sandstones, but they are less well-consolidated than those of the Anguille Group. They also contain layers of shale, coal, gypsum, rock salt, and other evaporites, which render the terrane as a whole weak with respect to erosional agents. Riley (1962) has mapped many folds and faults in these rocks that trend southwest to northeast. Dips in the folds are steep as is shown by coastal exposures, where beds commonly are nearly vertical. Streams draining across this lowland, from the Long Range towards the west, owe their present courses to their superposition on to a cover of glacial deposits, the thickness of which varies from a few feet near the mountain front to over 100 feet along the present coast.

The lowland now occupied by Deer Lake and the middle Humber valley, northeast of Corner Brook, is also underlain by weak Carboniferous sediments. In preglacial times a major river rose on a divide along the crest of the Long Range near Corner Brook and flowed to White Bay along this lowland. Drainage modifications in glacial times have reversed this flow, which now reaches the Bay of Islands through a breached divide in the lower Humber valley. This drainage change and others, will be discussed in Chapter Three.

The geologic and topographic variety of the bedrock areas of southwestern Newfoundland provided a diverse environment in which glacial

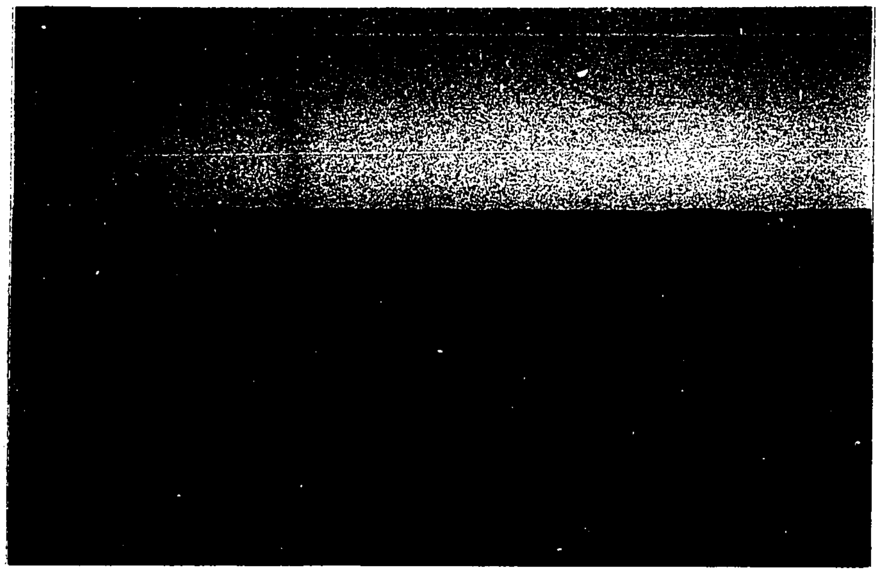

Figure 8: The smooth surface of the Cape Anguille Mountains. These valleys on the western flanks have been faintly modified by glaciation. Distant skyline of Long Range Mountains.

Figure 8: The smooth surface of the Cape Anguille Mountains. These valleys on the western flanks have been faintly modified by glaciation. Distant skyline of Long Range Mountains.

processes operated. Plateau surfaces of low relief were both etched by moving ice and covered with a thin and discontinuous mantle of irregular hummocky debris by stagnating ice. Smooth lowland areas provided paths through which piedmont glaciers flowed in late-glacial times. Isolated upland areas were the loci of ice masses during the decline of the island ice sheet and ice flow radiated from these. Major river valleys following structural depressions were overdeepened to form impressive canyons and fjords.

Such a variety of landscapes is atypical of Newfoundland as a whole. The common Newfoundland scene - a low "ria" coastline, with a rock-strewn land surface rising inland to boggy or forested plateaus of moderate elevation and low relief - is nowhere to be found along the western shore of the island. One reason is that the shoreline generally trends parallel to the strike of the rocks and structures. Another is the great elevation and relief which characterizes the area. The scenic beauty arising out of these factors is unmatched in the Atlantic Provinces.

CHAPTER TWO

A Review of Earlier Investigations of Glaciation in Newfoundland

The glaciation of Newfoundland has been investigated by several of the foremost figures in the field of Quaternary research. Investigations have focussed on two main problems. The first is that of glacier flow directions and the related problem of the relative dominance of an island ice cap and ice from Labrador. The second is that of late- and postglacial marine overlap and isostatic uplift. The two are intimately related and will be discussed together in this chapter, which should be read with Plates IV and V to hand.

The discussion is divided into two parts; the first dealing with the "pioneer" period of investigation, and the second with the "modern" period. In reviewing works published before 1940, one is struck by the scanty evidence upon which the weighty conclusions of authors were based. The main reason for this was probably the difficulty of mobility at that time in inhospitable and often uninspiring terrain. Before the confederation of the British overseas territory of Newfoundland with Canada in 1949, long-distance travel on the island was limited to the trans-island railway, which passes across some of the dullest terrain in the province, and to vessels plying the coastal waters on a more or less regular basis. The lack of accurate large-scale topographic map coverage of the island was an additional hindrance before the early 1950's.

Since confederation in 1949, the entire island has been covered by aerial photographic survey and 1:50,000, 1:250,000, 1:500,000, and 1:1

million topographic maps have been published. Detailed geological survey, begun by the pre-confederation Newfoundland government, has proceeded rapidly, so that several 1:250,000 preliminary and final maps are available. Further, the development of forest, mineral, and water-power resources, together with the increasing awareness of the importance of communications to economic and social development, has provided a network of paved and gravel roads.

Investigations undertaken since the late 1930's show a growing sophistication, with the gradual lessening of logistical restrictions. The refinement of the concepts of continental glaciation, and an increasing awareness that the classical glacial sequence, reconstructed in the American continental interior in the first three decades of the century, could not be applied in far distant areas of the continent, have provided a basis for the elucidation of events in small areas, but to date a comprehensive picture of Newfoundland glaciation has yet to emerge.

1. THE "PIONEER" PERIOD: pre-1940

The work of Bell (1884) was mainly concerned with the mainland of Labrador but, from a cursory examination of glacial features, he concluded that "on the island of Newfoundland, the glaciation appears to have been from the centre towards the sea on all sides" (p. 37DD). DeGeer (1892) produced a map of "The Last Changes of Level in Eastern North America" which showed isobases on the marine limit. Very few accurate determinations of marine limit elevations had been made at that time, but the map shows a zero isobase with a wide and deep re-entrant up the Gulf of St. Lawrence, and a corresponding southeastward bulge around the southern and eastern shores of Newfoundland. The 200-foot isobase is roughly parallel, passing eastwards through Anticosti Island and turning sharply northwards over central Newfoundland. The trend of these lines could be interpreted as evidence either of a southeastern lobe of Labrador ice or of an independent Newfoundland ice mass.

Chamberlin (1895) viewed the absence of far-travelled erratics and the presence of striae displaying a radial pattern on the Avalon Peninsula of eastern Newfoundland as evidence for "an extremely local derivation and a very short transportation of the drift, [and that] glaciation of the island was more probably attributable to the development of local ice sheets than to an extension of the ice fields of the mainland" (p. 407).

The map of Fairchild (1918), "Post-Glacial Continental Uplift" (his figure 1, p. 202), shows a well-defined dome of uplift over Newfoundland. Though the values on the isobases do not correspond with those of DeGeer, the southeastward bulges over the island in the zero, 100-foot, and 200-foot lines remain. Most of Fairchild's isobases over Newfoundland were hypothetical, with a few segments based on measured elevations. Some of these have subsequently been shown to be erroneous. For example, Daly's (1902) estimation of 575 feet at St. John's was used, but Daly corrected this figure in his paper of 1921. In spite of the lack of reliable evidence, Fairchild's conclusions are pertinent:

> All available information on the glacial geology of Newfoundland favours an independent ice-cap, with radial flow. And when we consider the large area and mountain heights of the island; the northern position; exposure to ocean on all sides; and location in the paths of the cyclonic storms of America, it appears certain that the island was the locus of heavy snow precipitation and a massive ice-cap - the Newfoundland continental glacier. (p. 229.)

J. B. Tyrell, writing of a visit to Newfoundland in 1917, favoured an independent Newfoundland ice cap:

> In Newfoundland the last glaciation, which was very strong, was from the center outward both eastward and westward. If the glacier from Labrador had ever reached the island, evidence of its presence would have been discernible on the west coast, but in the short time I was there I was not able to find any such evidence. (In Fairchild, 1918, pp. 227-228.)

Daly (1921), after revising a few of the elevations of post-glacial marine features which he determined in 1902, reached a related conclusion: "It is possible that Newfoundland has been faintly domed or arched and that the final mapping of the isobases will indicate the secondary influence of the independent ice-cap of Newfoundland on the character of post-glacial warping" (p. 385).

Glacial striae displaying a radial pattern in Daly's map (see Plate V) were used as supporting evidence for the above statement. The striae, however, are mainly in coastal valley sites, where topographic control of flow would be strong in the later stages of glaciation. They say little of the presence of a Newfoundland ice cap at the glacial maximum.

Coleman (1926), following a brief study of glacial features and drift provenances on the island, concluded that Newfoundland had supported its own ice cap before Wisconsin time[1], and that ice flow radiated outwards to beyond the present shoreline, covering all but the highest summits on

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1. Coleman's assignment of marine deposits in the Bay of Islands (Plate I) to an "interglacial" forced the assignment of underlying till to pre-Wisconsin time. The marine deposits are now known to be late-glacial in age and hence the underlying till is presumably main Wisconsin.

the west coast. He believed that the "Wisconsin" [sic] glacial episode was multi-centred, and many summits in the central part of the island stood above the ice sheet. Coleman's conclusion that the highest summits on the west coast had not been glaciated was supported by Fernald (1925), who saw those summits as glacial refugia for certain plant species of circum-Arctic and Cordilleran affinities. Much debate has centred around the distribution of these plant species in mountain areas in boreal North America and Scandinavia (Wynne-Edwards, 1937; Dahl, 1955) and early conclusions have been discredited in some areas. The hypothesis is linked to the more recent "nunatak hypothesis" which calls for ice-free summit areas in boreal mountain regions to explain the presence of high-level block-fields. Dahl (1966) and Ives (1966) have recently stated opposing views on this hypothesis.

The works of Bell, DeGeer, Chamberlin, Daly, Fairchild, Tyrell, and Coleman belong to the "pioneer" period of glacial research in Newfoundland. The common element in all their conclusions is the belief that the island of Newfoundland supported its own ice cap at the last glacial maximum. No hint of the invasion of Labrador ice is found.

2. THE "MODERN" PERIOD: post-1940

In 1940, the Geological Society of America published papers by Flint, and by MacClintock and Twenhofel. This work was the result of investigations carried out in the late-1930's from Princeton and Yale universities, which were continuing an association with geological research in Newfoundland that existed through the thirties and forties.

Flint (1940) made a detailed investigation of raised marine features along the west coast of Newfoundland and included data from the other coasts and from the Maritime Provinces in his general conclusions. He recognized a group of "higher marine features" along the west coast of the island which were formed when ice from an island outflow centre wasted inland of the present shoreline. Many of these features were interpreted as having been formed in intimate association with active glacier ice. Flint reconstructed tentative isobases on the upper marine features and the resultant plane rises up towards the north-northwest. The lowest isobase is the 100-foot, which passes through the head of St. George's Bay (Plate V), and: "southward projection of this assumed surface would carry it down to present sea level near Port aux Basques" (p. 1771). The absence of raised marine features in the southwest corner of the island was attributed to the presence of late ice which delayed marine overlap past the time when isostatic rebound was great enough to have elevated shoreline features above the eustatically rising sea.

Flint tentatively reconstructed a lower tilted sea level plane, sloping up in the same direction as the higher one, from a zero isobase passing through Stephenville (Plate V) to 250 feet at the northern end of the Great Northern Peninsula. Many more features were available to assist in this reconstruction than the small number of features observed at the marine limit. Some rock-cut platforms which Flint included on this lower "Bay of Islands surface" prompted him to conclude that a stillstand had occurred which lasted for about 5,000 years. A stillstand of similar duration was postulated to account for shore platforms at present sea level.

> . . . the pronounced northwestward rise of the marine features on the west coast implies that the former glacial mass was thickening toward the northwest. From

> this it may be regarded as likely that at least western Newfoundland was invaded by the Labrador ice sheet, flowing from the northwest, and that local Newfoundland glaciers were not massive enough to interfere seriously with the regional crustal influence of the Labrador ice. (p. 1775.)

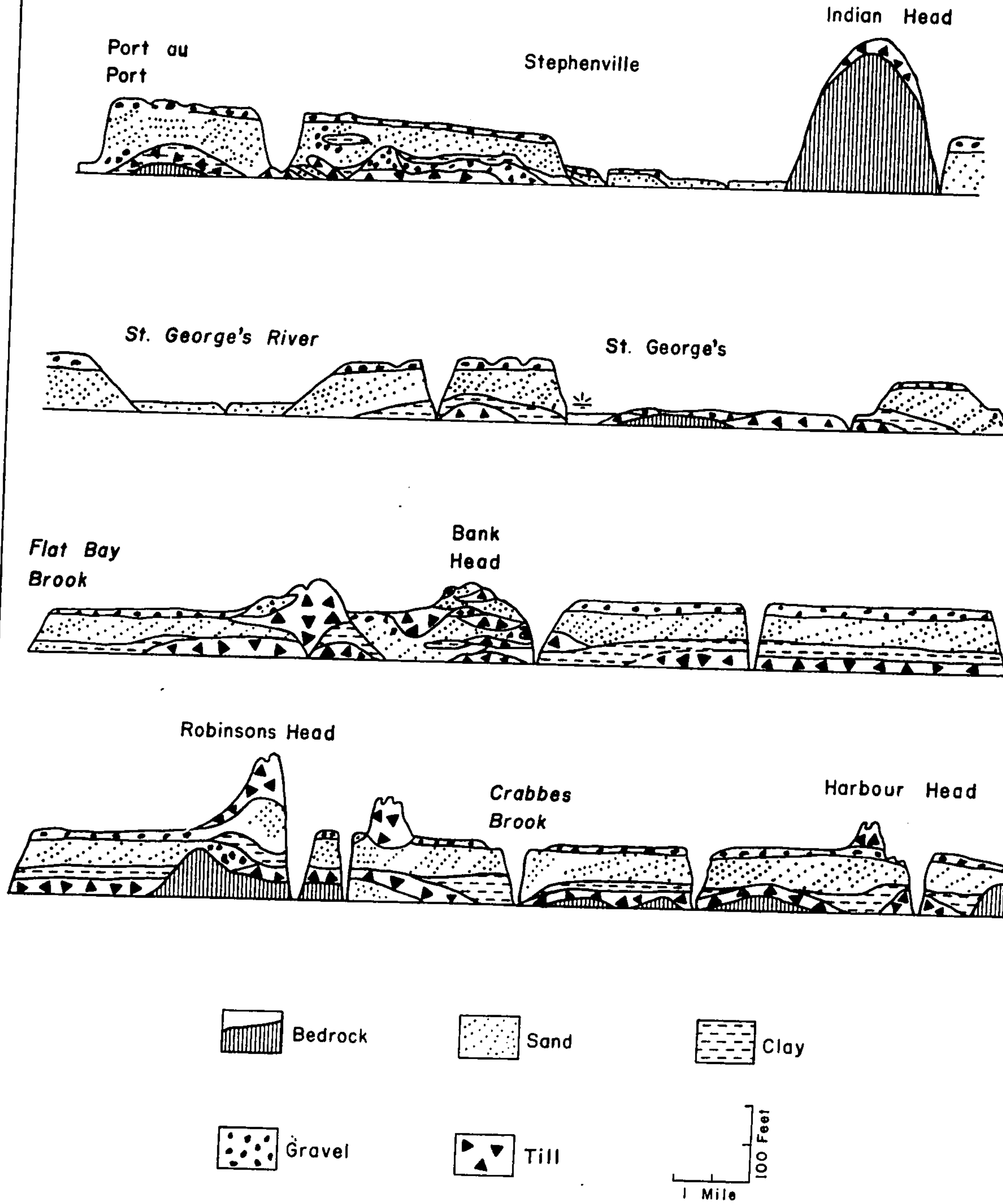
Since Flint ascribed the glacial event which preceded marine overlap to an island ice-cap stage, it can be interpreted that he viewed this as a relatively late re-orientation of ice flow directions which had been dominated by the invasion of Labrador ice at the glacial maximum.

The work of MacClintock and Twenhofel (1940) was the first to deal with Quaternary stratigraphy in Newfoundland. Their investigation was centred on the coastline and hinterland of St. George's Bay, but local studies were possible in isolated areas in other parts of the island. They found no evidence of a glaciation prior to the Wisconsin and none to support the earlier conclusion that high summits had stood above the ice at the glacial maximum.

In the St. George's Bay area, three Wisconsin lithologic units were recognized and were assigned to three contrasted events in the glacial history (figure nine). The lowermost unit comprised till and associated ice-contact deposits that had a limited surface exposure, but which were well exposed near the base of coastal cliffs. They gave the unit the name "St. George's River Drift." Overlying this unit was a sequence of marine deposits. The lowest of these were silty clays and clays with occasionally abundant marine shells. Delta bottomset and foreset and, in places, topset beds overlay them, often continuing to the top of coastal cliffs. Inland of the cliff edge, the level surface of the sediments was interpreted as an original glacio-marine delta. This unit was called the

Figure 9.

STRATIGRAPHY OF QUATERNARY DEPOSITS IN COASTAL CLIFF SECTIONS, ST. GEORGE'S BAY



after MacClintock & Twenhofel, 1940.

"Bay St. George Delta." MacClintock and Twenhofel gave the name "Robinsons Head Drift" to a sheet of glacial deposits spread widely over the St. George's Bay hinterland. The drift was composed predominantly of ablation moraine and its outer margin was marked by a prominent kame-and-kettle end moraine (Plate IV). This was seen to be cut by coastal cliffs in several places where stratigraphic relations could be seen between it and underlying and adjacent marine sediments. MacClintock and Twenhofel, and Flint (1940), recognized that the ice which produced the "Robinsons Head Drift" moved seawards over a gently shelving sea floor.

Later episodes in the glacial history of Newfoundland were recognized in small, looped end moraines near the mouths of canyons dissecting the Long Range Mountain front, and in cirques in these and other mountains.

MacClintock and Twenhofel (1940) reported glacial striae on bedrock ledges in the Port au Port area that were interpreted as recording ice movement first from the north and later from the east. The striae were associated with the "St. George's River Drift" stage.

> This succession of events would fit well within the proposal . . . that Newfoundland was at first affected by ice from Labrador and later had its own outward-radiating ice cap. . . . The ice which moved southward may have been either Labrador ice or Newfoundland ice which, farther north, moved out into the St. Lawrence Gulf, there encountered south-moving Labrador ice, and was pushed southwards by it. (p. 1753.)

These speculations on the origin of ice flowing southwards over the Port au Port Peninsula attempted to bring into harmony the conclusion of Flint (1940), quoted above, with their observation that "striae in most places indicate ice radiating outward from Newfoundland" (p. 1750). MacClintock and Twenhofel made the following general statement suggesting the sequence of events:

> glaciation from Labrador took place at the climax of the Wisconsin stage, when ice may well have passed not only over Newfoundland but out to deposit the well-known drift of the Grand Banks. . . . As the Labradorian glacier waned part of it may have persisted on Newfoundland, acquired radial motion as impeding ice melted away from its margins, and thus became a local ice-cap whose movement would have largely obliterated any earlier striae. (p. 1750.)

By 1940, then, the earlier concept of a dominant Newfoundland ice cap had been replaced by that of overriding Labradorean ice at the glacial maximum, followed by a late-glacial highland centre of radial outflow.

Murray (1955), working in the level area of the central plateau of Newfoundland, reported striae that he interpreted as due to ice moving from north to south towards the south coast of the island. He offered two alternative explanations of these. The first deferred to Flint's earlier conclusion that Labrador ice had invaded at least western Newfoundland (and tentatively extended by MacClintock and Twenhofel to cover the whole island). Southward flow over south-central Newfoundland can be envisaged as a consequence of this. The second relates the striae to relatively late-glacial conditions in which a steeper land surface slope to the south from the Annieopsquotch Mountains caused ice to flow more rapidly in that direction. The latter explanation is the more likely one and can be supported by the suggestion that late-glacial calving of the ice front into marine waters along the south coast of the island would have rapidly steepened the slope of the glacier and caused it to flow more rapidly southwards.

Jenness (1960) has contributed the only comprehensive work on the glaciation of eastern Newfoundland. Plate IV shows the main features of glaciation reported in his study. Jenness recognized an "outer drift zone" in which deposits are "the products of advancing ice . . . that is . . . ground moraine, largely of local origin, . . . end moraines, drift ridges, eskers and kames, kame terraces, and large patches of glacial outwash trending seaward from the margins of the inner drift zone" (p. 168). Only small occurrences of end moraine were identified, so it is hazardous to draw conclusions from them as to the extent of the ice mass at the time of deposition of the "outer drift zone." Jenness stated, however, that the ice margin lay offshore in most places.

Separating this "outer drift zone" from an "inner drift zone," Jenness traced a discontinuous end moraine (Plate IV) which represented either a still-stand in the back-wasting of the ice mass or a readvance from farther inland. At this stand, streams issuing from the ice front built outwash valley trains and deltas graded to a sea level higher than present.

An area of fluted terrain in the "inner drift zone" is worthy of discussion in this context (Plate IV). Jenness concluded that this was a product of the last of the glacial stages to affect that area; presumably the stage at which the end moraine was built. The fluted terrain lies astride the height of land north of Fortune Bay, and features on it are oriented northwest to southeast on both sides of the height of land. This suggests that the ice that produced the flutings was so thick that flow from the northwest was not influenced by the land slope on each side of the divide. This would seem to militate against the late-glacial origin for the feature. It is to be expected that in late glacial times ice would be thin enough to have had its basal flow controlled by land surface slopes. It is significant that, in the area of the fluted terrain, striations are oriented northwest to southeast on both sides of the west-east-trending divide, but eskers closely follow the land surface slopes away from it. This indicates that active ice moved from the northwest over this divide, and that the eskers, associated with the "inner drift zone" were later deposited beneath a thinner, fissured, inactive ice mass. The active ice moving southeastward could have originated in the Labrador centre, but Jenness preferred to call upon ice from "the western half of the island" (p. 163).

Jenness constructed isobases on the highest marine features recorded on the northeast, east, and southeast coasts of Newfoundland. These are shown in Plate V. Most of the features are delta terraces which were built into the sea when the ice sheet had receded to the position of the moraine separating the outer and inner drift zones. They are probably contemporaneous and the isobase reconstruction is, therefore, probably valid. Jenness tentatively projected these isobases to join those which Flint (1940) drew on west coast features. The curvature of the resultant isobase pattern, showing a southeastward bulge similar to that of Fairchild and deGeer, prompted Jenness to keep open the door to Labrador ice by stating that the isobase trend "cannot but promote renewed speculation that Newfoundland was invaded by the Labrador ice sheet" (p. 177). This curvature, however, could also reflect the influence of a Newfoundland island ice dome. Radiocarbon dates on marine shells in deposits relating to the marine limit on the northeast coast have become available since

Jenness' work. These show that marine overlap occurred approximately 12,000 years B.P. (Dyck and Fyles, 1963). The position of the dated localities is shown on Plate V. The dates invalidate Jenness' tentative extension of east coast isobases to those drawn by Flint (1940) on the west coast. Whether or not the latter are correct, the features upon which they are based are older by up to 1,500 years than those on the east coast.

The work of Lundquist (1965), in the area inland of Notre Dame Bay, produced results similar to those of Jenness (1960) with regard to glacial landforms. On the evidence of intersecting sets of glacial striae and flow-moulded landforms, Lundquist concluded that the high plateau areas of the Long Range Mountains had acted as glacial outflow centres at a stage of the island ice cap. Ice moved towards the northeast beyond the present shoreline and subsequently retreated inland to permit the overlap of marine waters. During the latter phase the ice margin halted temporarily along a line presently indicated by a junction between thick ground moraine inland and a thinner till cover towards the coast. Lundquist called this a "supposed line of retardation" (in ice retreat), and it conforms closely to the position of the end moraine that Jenness mapped between his inner and outer drift zones.

During further wastage of the glacier, it became increasingly influenced by local topography, as evidenced by striae trending northwest to southeast on interfluves inland of Notre Dame Bay. Some of these striae Lundquist saw as possible evidence that "the continental ice sheet from Labrador at an early stage - probably early Wisconsin - affected Newfoundland, possibly the entire main part of the island" (p. 304).

3. SUMMARY

This review has demonstrated a continuing concern with (a) the relative influence of Labradorean and "island" ice on the glacial deposits landforms and post-glacial uplift and (b) the nature of marine overlap. Table I is a summary of previous works on the glaciation of Newfoundland. Early opinions that an independent ice cap lay over Newfoundland were based on very tenuous evidence. Flint's conclusion that Labrador ice invaded the west coast at the glacial maximum, and the tentative extension of that conclusion by MacClintock and Twenhofel to the entire island, initiated a period of entrenchment of that idea in the literature and in the minds of others whose attention has been drawn to the glaciation of Newfoundland. In the present study the evidence for an "island" ice cap at the Wisconsin maximum is discussed in Chapter Four.

TABLE I

SUMMARY OF PREVIOUS INVESTIGATIONS OF GLACIATION IN NEWFOUNDLAND

AUTHOR(S) AND DATE	STUDY AREA	AGE OF GLACIAL STAGES	AREAL EXTENT OF GLACIATION	DIRECTION OF GLACIER FLOW	VERTICAL EXTENT OF GLACIATION
Chamberlin 1895	Avalon Peninsula	No comment	Local ice cap	Radially seawards	Complete cover
Tyrell 1917	Western Newfoundland	"the last glaciation"	No comment	Radially seawards	No comment
Daly 1921	Newfoundland	"Wisconsin"	Beyond present shoreline	Radially seawards	No comment
Coleman 1926	Newfoundland	1) Kansan or Jerseyan 2) Interglacial marine episode	Beyond present shoreline	Radially seawards	1) Nunataks in western highlands
		3) "classical" Wisconsin	No comment	Radially seawards from minor centres	2) Nunataks in western and central highlands
MacClintock and Twenhofel 1940	Newfoundland, mainly west	"Wisconsin"	1) Beyond present shoreline:	Radially seawards	Complete cover

TABLE I (Continued)

AUTHOR(S) AND DATE	STUDY AREA	AGE OF GLACIAL STAGES	AREAL EXTENT OF GLACIATION	DIRECTION OF GLACIER FLOW	VERTICAL EXTENT OF GLACIATION
MacClintock and Twenhofel 1940			"St. George's River Drift stage"		
			2) Marine overlap on west and northeast coasts: "Bay St. George Delta stage"		
			3) Readvance to St. George's Bay shoreline: "Robinsons Head Drift stage"	Seaward, around St. George's Bay	No comment
			4) Local ice caps: "Kittys Brook moraine stage"	Radially from minor centres	No comment; summits probably ice-free
			5) Local valley and cirque glaciation	Down-valley	Summits ice-free
Jenness 1960	Eastern Newfoundland	Late Pleistocene	1) Island ice cap, margins beyond present shoreline, with separate, contiguous cap over Avalon Peninsula	Radially seawards, stronger from northwest	No comment
			2) Marine overlap		

TABLE I (Continued)

AUTHOR(S) AND DATE	STUDY AREA	AGE OF GLACIAL STAGES	AREAL EXTENT OF GLACIATION	DIRECTION OF GLACIER FLOW	VERTICAL EXTENT OF GLACIATION
Jenness 1960			3) Island ice cap, end moraine inland of shoreline	Radially from minor centres	No comment
Geological Survey of Canada (Henderson, in Dyck and Fyles, 1963)	Northeastern Newfoundland	12,000 years B.P.	Island ice cap clearing present shoreline, followed by marine overlap	No comment	No comment
Lundquist 1965	Northeastern Newfoundland	Wisconsin	1) Probably early Wisconsin Labrador ice cap	Northwest to southeast	No comment
			2) Later Wisconsin island ice cap	Southwest to northeast beyond present shoreline	Central plateau monadnocks above 1,500 feet ice-free
			3) Marine overlap		
			4) Island ice cap, stillstand inland of shoreline	Southwest to northeast beyond present shoreline	No comment
			5) Minor centres, in lowlands	Radially from centres	Central plateau summits emerge from ice

CHAPTER THREE

Some Features of Glacial Erosion in Southwestern Newfoundland

Because this study is mainly a regional stratigraphic-chronologic one, no attempt was made to investigate the landforms of glacial erosion in a comprehensive way. However, aspects of these are so prominent in the landscape of some parts of southwestern Newfoundland that to omit them would detract from the study rather than enhance it. It is for this reason that the erosional forms investigated in the process of studying the stratigraphic aspects of glaciation are considered in this chapter. The treatment does not purport to be complete. Rather, it attempts to shed light on the origin of the observed features and to draw some conclusions from them about glaciological conditions at the time of their formation.

1. SMALL-SCALE FEATURES

In a recent comprehensive treatment of glacial and periglacial geomorphology, Embleton and King (1968) have classified small-scale features of glacial erosion on bedrock, as follows:

Striae	plastic scouring
grooves	plucked surfaces
polished surfaces	"roches moutonnées"
friction cracks	(stoss-and-lee topography)

All of these features are found, more or less well developed, in southwestern Newfoundland. Where they are best developed, conditions responsible for their formation and preservation have been optimal. The glacial striae have been studied by MacClintock and Twenhofel (1940) for the evidence that they provide of ice-movement directions at different times during the glaciation of the area. The other features have received no attention beyond the casual remarking on their presence in some localities. In this study, all the small-scale features of glacial erosion, except for the plucked surfaces and "roches moutonnées," have been investigated in the area of the Port au Port Peninsula, where they are presently seen on smooth ledges of Ordovician limestone emerging from beneath a cover of the lowermost glacial deposits, the "St. George's River Drift" of MacClintock and Twenhofel (1940). Plucked surfaces and "roches moutonnées" were only cursorily examined in the Long Range Mountains east of St. George's, where they are prominent on an anorthosite outcrop. In a few other localities, such as the sides of fjords in the Bay of Islands and Bonne Bay, striae were studied solely for evidence of ice-movement directions. The discussion of small-scale features of glacial erosion, then, is limited to those forms seen in the Port au Port area on limestone bedrock. Sufficient diversity was recorded to warrant an attempt to explain the forms there.

Along the southern shore of the Port au Port Peninsula and around the shores of East Bay, Port au Port Bay, ledges of Table Head limestone are overlain by St. George's River Drift till. The surface of the grey, massive limestone, within 25 feet of sea level, is glacially eroded and many of the small-scale features of glacial erosion are seen upon it. That many of these features, particularly striae, occur close to sea level may arouse the suspicion that sea-ice might have played a rôle in their formation. However, four factors may be cited in favour of a glacial origin. Firstly, while many localities are close to sea level, some are up to 25 feet above sea level, and therefore above the influence of sea-ice for several thousands of years before present. Secondly, apart from striae, grooves, friction cracks, plastic scouring features, and miniature "rock drumlin" forms are also seen (figures 10, 11, and 12). Such features cannot be formed by sea-ice and, furthermore, were produced by an erosive agent moving towards the sea in many cases. Thirdly, the evidence provided by these features of the direction of movement of the erosive agent corroborates that provided by the provenance of erratic boulders in the St. George's River Drift till. Fourthly, in all localities where the micro-features of glacial erosion are found, they can be traced to the present seaward edge of the till overlying the bedrock surface and disappear beneath the till. If the till is cut away from the bedrock surface, the erosional features can be seen to continue across the latter. An origin by any agency other than flowing glacier ice is, therefore, ruled out.

The high density and relative purity of the Table Head limestone, together with the fact that, in the Port au Port area, the dip of the strata is everywhere low, has rendered the bedding planes of the rock susceptible to smoothing, polishing and small-scale erosion. The fact that where the erosional features are visible they have only recently been exposed by wave and spray erosion of the superincumbent glacial till, makes for an excellent state of preservation at the present time.

The nature of the St. George's River Drift till is significant for an understanding of the erosive agent responsible for the features. Mechanical analysis, presented in figure 22, Chapter Four, show the till to be relatively rich in clay and silt. Clasts in the till are predominantly igneous and metamorphic fragments with an angular and subangular shape. Such material, rich in fines and including many strong, small

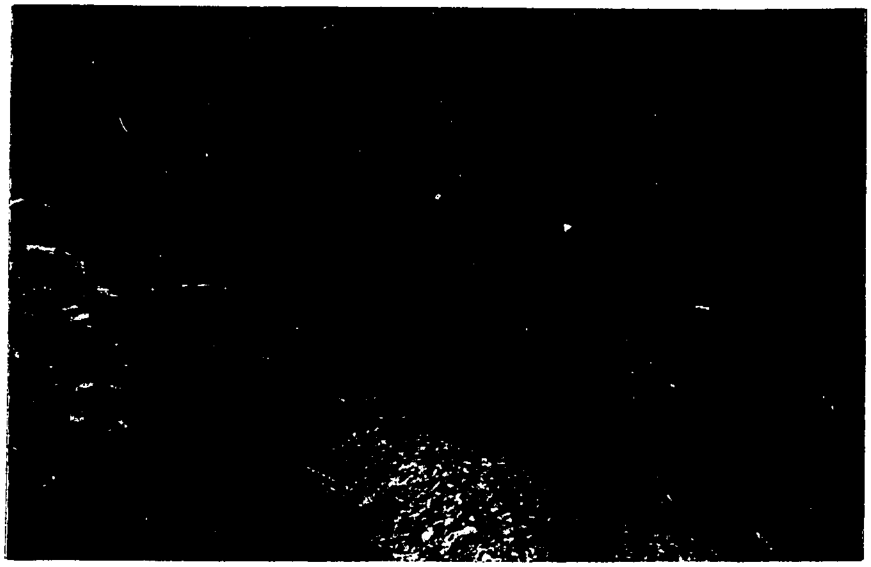


Figure 10: Glacial fluting and miniature rock drumlins, Sheaves Cove, Port au Port Peninsula. Ice flowed towards observer. Pen gives scale.

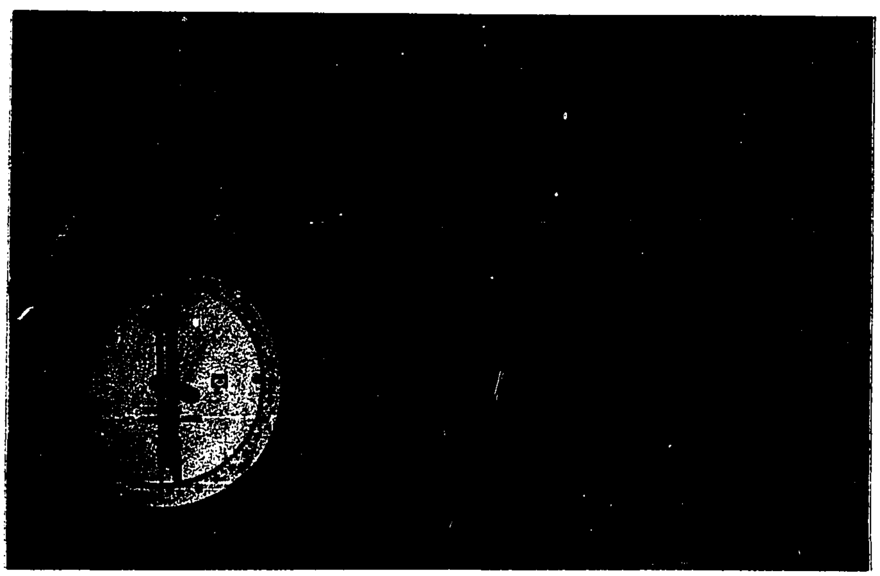


Figure 11: Crossing glacial striae at Campbells Creek, Port au Port Peninsula, indicate strong ice flow first from 052° True and later weaker flow from 000° True. Magnetic declination 28°W.

Figure 10: Glacial fluting and miniature rock drumlins, Sheaves Cove, Port au Port Peninsula. Ice flowed towards observer. Pen gives scale.

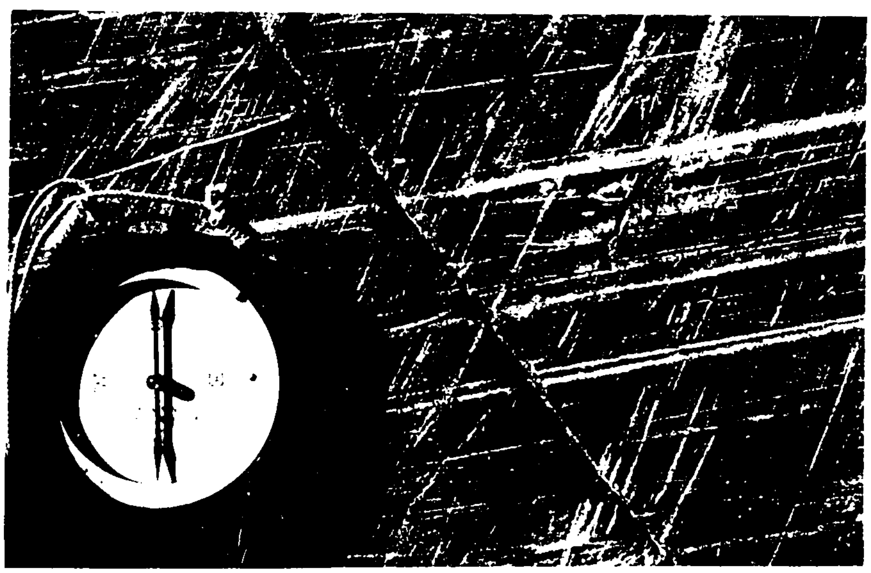


Figure 11: Crossing glacial striae at Campbells Creek, Port au Port Peninsula, indicate strong ice flow first from 052° True and later weaker flow from 000° True. Magnetic declination 28°W.

clasts, would be an extremely effective erosive agent, especially in an unconsolidated "sludge" condition, such as Gjessing (1965) viewed as important for "plastic scouring." Gjessing envisages a basal debris-water-ice mixture as the main agent in this process.

The regional trends of glacial striae recorded in the Port au Port area follow three compass directions (see figure 23, Chapter Four). The earliest striae were produced by ice flowing westwards from the highland centre of outflow over the Long Range Mountains. Later striae follow the general trends from bearings up to 25 degrees east and west of north. The earlier striae are deeper and wider than the later sets and the latter do not always cross the former. Where this latter relationship occurs, it is easy to be led to the conclusion that the east-west striae are the later set. However, the relationship may better be explained as the result of a momentary release of pressure, and therefore striating power, at the base of ice passing from northerly points across the deep east-west striae. The temporal sequence is properly established when other glacial erosion features are considered, and more will be said of this in Chapter Four.

"Crescentic gouges" as defined by Harris (1943) are common in the Port au Port area (figure 11) and are exclusively associated with the striae which record ice-movement from easterly points. Often the gentle slope between the horns of a gouge is itself striated in a direction related to flow from northerly points. This latter flow, then, is the later.

Grooves are found across all the limestone surfaces where other features are seen. They occur on several scales, ranging from wavelengths transverse to the direction of ice-movement of 3 to 4 feet to a few inches. At Campbells Cove, one example of a groove one inch deep, was seen with

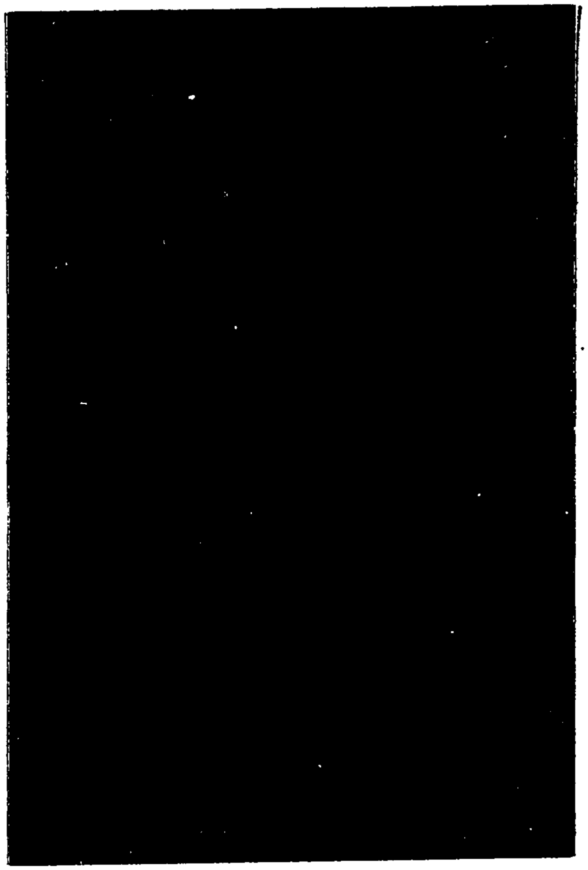


Figure 12: Crescentic gouges and friction cracks on limestone surface, Abrahams Cove, Port au Port Peninsula. Pen points in direction of ice flow.

Figure 13: Fjords of Bonne Bay. Maximum depth of 126 fms. in East Arm, out of view at right.

Figure 12: Crescentic gouges and friction cracks on limestone surface, Abrahams Cove, Port au Port Peninsula. Pen points in direction of ice flow.

Figure 13: Fjords of Bonne Bay. Maximum depth of 126 fms. in East Arm, out of view at right.

right-angled turns to the right, then left, in its 15-inch length. Miniature "rock drumlins" one half to one inch across and from one to several inches in length occur on the sides of minor inclusions of less pure limestone in the direction of ice flow. These grooves and "rock drumlins" attest to the plasticity of the erosive agent and also to its strength (figure 10). It is not difficult to envisage that such a plastic medium could be responsible for such powerful erosion if it is considered that the stress would have been applied over relatively long periods of time.

In only one case were small-scale erosional features recorded which could not be explained by the scouring action of a sub-glacial, plastic, debris-ice-water mixture. This is an area of friction cracks which look superficially like the crescentic gouges seen at many localities in the Port au Port area. At Abrahams Cove, on a limestone bedrock ledge which has been glacially grooved and smoothed, a set of straight cracks, two to three inches long, traverse a crest between two grooves perpendicular to the trend of the grooves (figure 11). No chips of rock have been removed from between these cracks. Several inches away, in the direction towards which ice flowed, lunate chips have been removed from a series of several cracks which are left with a vertical face on the "up-ice" side of the crack and a face sloping gently upwards in the "down-ice" direction between "horns" which point "up-ice." The fracture along which the chips of rock have been removed dips into the rock in the direction from which ice flowed. This is opposite to the dip of that crack in the case of other "crescentic gouges."

Preston (1921, cited in MacClintock, 1953), suggested, from the evidence of experiments with metal balls moved under pressure across glass surfaces, that a crack which dips in the direction of movement of the force producing it results from a fragment being rolled over the surface in the same direction. A crack which dips in a direction opposite to that of the force is produced by a fragment sliding across the surface under pressure.

These transverse cracks and crescentic gouges at Abrahams Cove, then, probably were produced by ice sliding rapidly over the bedrock surface so that clasts in the base of the ice were not rolled. This may have been due to the locally steep ice gradient produced by channelled flow in a preglacial bedrock valley. The form of this valley can be well seen from seaward as the surface of the bedrock exposed in the coastal cliffs drops steeply below sea level at the sides of the cove.

These erosional micro-features shed some light upon the glaciological conditions which were likely to have prevailed during their formation. The plastic scouring forms seen on the limestone bedrock of the Port au Port shoreline attest to the presence of a basal ice-water-debris mixture which was yielded into a plastic condition by the stress of superincumbent ice. Even at the maximum of Wisconsin glaciation the Newfoundland ice cap would have been a "temperate" glacier, with temperatures near the pressure-melting point of ice throughout its thickness. Manley (1955) concluded that the firn limit would have lain near present sea level at the Wisconsin maximum in southeastern Canada, therefore mean annual air temperatures would have been sufficiently high to inhibit great refrigeration of the glacier.

Under these conditions, ice flowing from highland outflow centres in the mountains of western Newfoundland into the Gulf of St. Lawrence would have considerable erosive ability. Most of the strong erosion by the ice sheet appears to have taken place, however, along the steep margins of those mountains and across the relatively weak Carboniferous sediments east of St. George's Bay. Upland plateaus in the Long Range Mountains appear only to have been weakly etched by ice, whereas the steep western margin of the range shows much evidence of rock removal, particularly in the major valleys through which channelled flow would have occurred. West of the mountains the Carboniferous sedimentary rocks were susceptible to glacial erosion by virtue not only of their mechanical weakness but also because they strike predominantly at right angles to the direction of ice flow.

The weakness of glacial striae which were formed by ice moving from northerly points, at a later time than the main westward movement, indicates a weakening of the erosive ability of ice at the time immediately prior to marine overlap. It was only when marine waters encroaching into the Gulf of St. Lawrence had caused the ice margin to be oriented parallel to present shorelines that flow in the Newfoundland ice cap changed from westward to southward over the Port au Port Peninsula. Forward movement continued, forming weak striae and transporting erratic boulders at least 30 miles from the north and northeast. No features indicative of more vigorous flow are seen associated with these striae as they are with the earlier east-to-west set. The ice sheet can, then, be envisaged as thinner and less well nourished than at the stage of the main island ice cap.

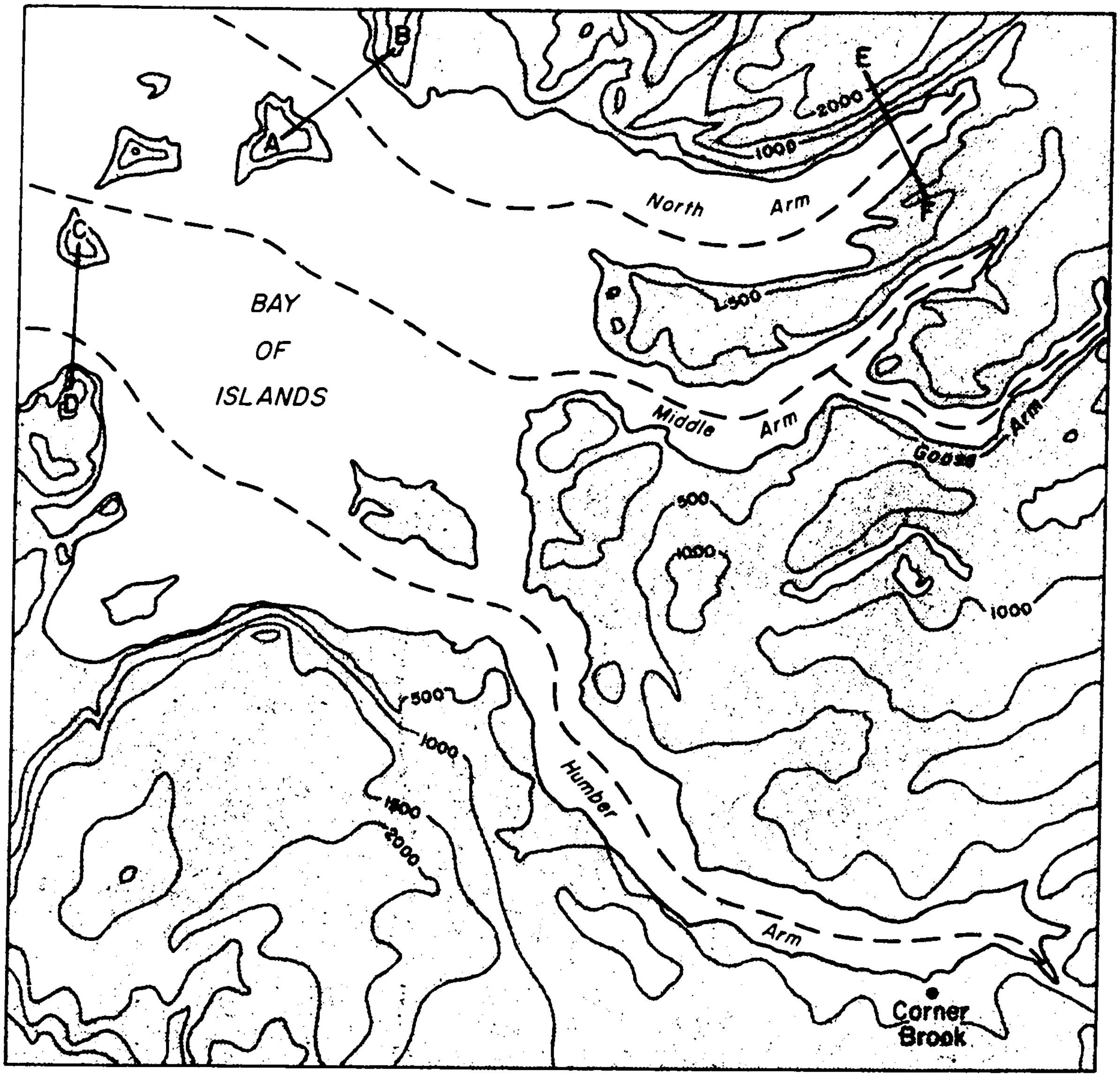
Further discussion of ice-cap dimensions and the glaciation at the Wisconsin maximum is delayed until Chapter Four, after the St. George's River Drift deposits have been dealt with.

2. FJORDS

The steep margins of fault-bounded plateaus in western Newfoundland, and the valleys cut into less resistant rocks between them, provided optimum conditions for strong erosion by valley glaciers. The fjords of the Bay of Islands and Bonne Bay are among the most spectacular results of glacial erosion in the area. The interest of the writer became focussed on the fjords rather than on the glacial troughs further from the sea because of the implications of any conclusions reached regarding the period of marine overlap for the mode of origin of the fjords. Most of the attention was paid to the Bay of Islands fjords because, not only are detailed marine charts available which permit accurate profiles to be drawn of them, but a fairly detailed picture of the late-glacial history of that Bay has emerged from the study, which enables any theory of fjord formation to be assessed. Soundings are also available for Bonne Bay (figure 13) and for St. Paul's Inlet, at the southern end of the Great Northern Peninsula, which identify them as fjords. A fjord is defined here as a steep-walled, glacial trough that has been eroded to depths well below present sea level, with one or more deep basins set into the general long-profile, and with a seaward threshold between the outer-most basin and the offshore zone.

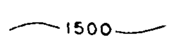
This form is well displayed by the Bay of Islands fjords, shown in figure 14. Profiles are shown in Plate VI. Crary (1966) has proposed a mechanism of fjord formation which has particular merit in these fjords, and, by analogy, in the others found in western Newfoundland.

Figure 14.



THE BAY OF ISLANDS

LOCATION OF FJORD LONG PROFILES

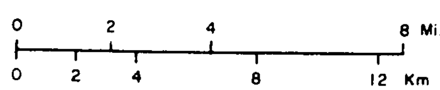


Contours on land areas at 500 foot intervals

A ——— B Lines of cross profiles shown on End Plate

— — — — Lines of long profiles shown on End Plate

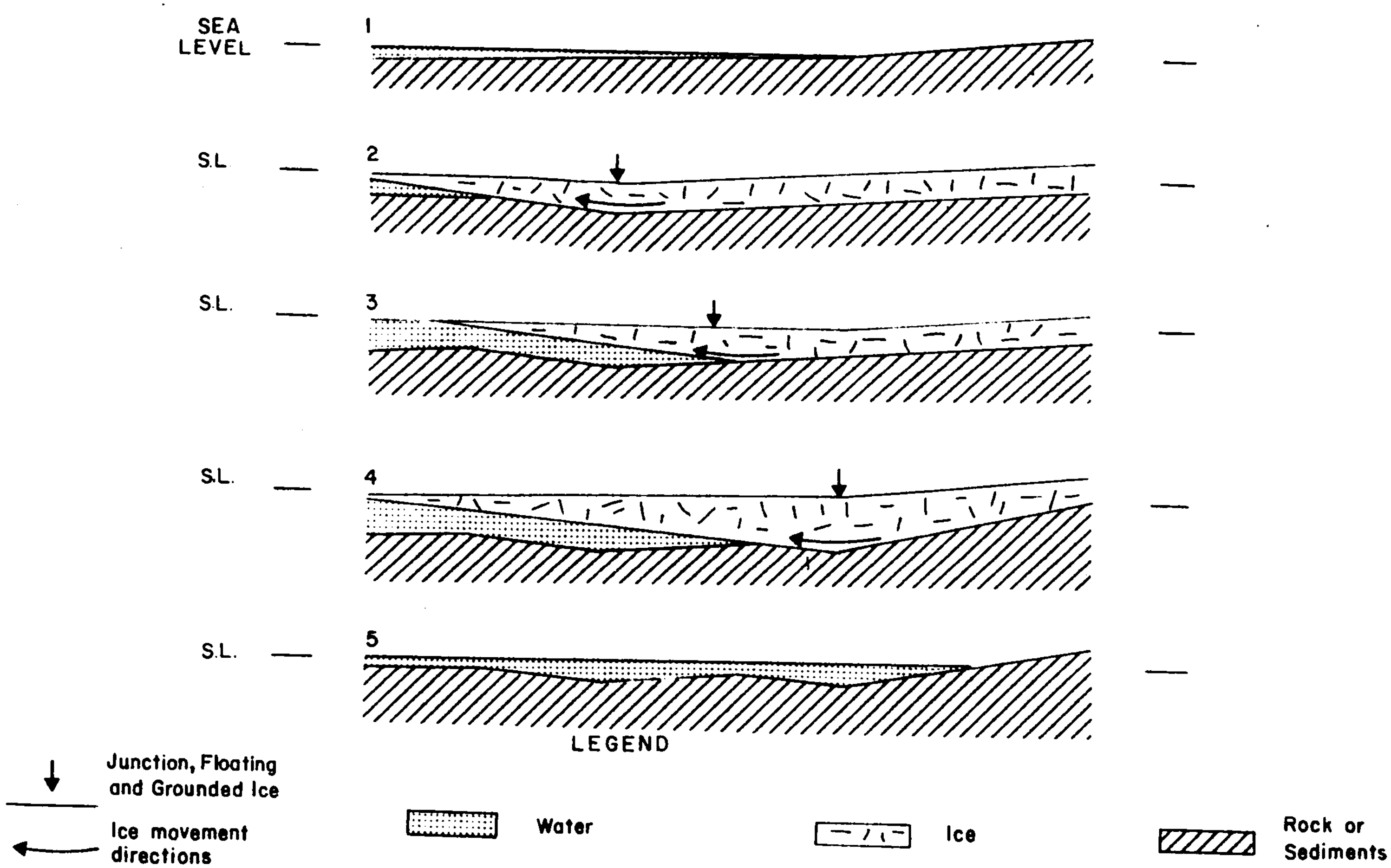
(See Plate VI for profiles)



In a fjord presently forming at the inland edge of the Ross Ice Shelf in Antarctica, seismic profiles show a glacier tongue descending a valley cut in crystalline rocks and grounding on the floor of an inlet, in a basin 1,500 metres below sea level. Seaward of the basin the glacier tongue joins a floating ice shelf. The ice thickness increases from 500 metres at the inner edge of the shelf to 1,500 metres at the head of the inlet, an increase which is attributed by Crary to the decreasing width of the valley. Crary saw the deepening of the fjord floor just inland of the contact of glacier and shelf ice as due to the scooping activity of glacier ice at a place where ice movement is mostly towards the terminus of the glacier (figure 15). He draws an analogy with an "endless belt or chainsaw . . . any increase in thickness or uplift of the land would bring the bottom material gradually in contact with the moving underside of the ice, allowing direct erosive action to take place" (p. 927).

There are several implications to this theory. Firstly, rock basins in fjords are eroded earlier to seaward and decrease in age inland as the sea penetrates an inlet. Secondly, the erosion of rock basins in the fjord floor does not begin until the sea penetrates beneath a glacier tongue to float its distal portion. Thirdly, in areas where rapid late-glacial isostatic uplift was occurring immediately prior to deglaciation of the fjord valleys, there was a relatively short period of time available for rock basin cutting, since sea level would quickly be drawn down possibly below the level of the seaward threshold. Where late-glacial emergence did not raise the threshold out of the sea, rock basins could still be cut. Since the threshold of the Bay of Islands is 80 feet below sea

Figure 15. STAGES IN THE FORMATION OF MULTIPLE BASINS IN FJORDS (after Crary, 1966)



level and emergence there was less than this, rock basin cutting could have proceeded while there were glacier tongues in the valleys now occupied by the tide-water "arms." Fourthly, the cutting of the rock basin furthest to seaward would not have been possible until the stage when a glacier tongue in the valley terminated closest to the present mouth of the inlet, since the threshold separating it from the gently shelving sea floor off-shore lies across that mouth.

A radiocarbon date of 12,600 ± 170 years B.P. (G.S.C.-868) has been obtained for the marine limit at Cox's Cove, Middle Arm, Bay of Islands and a comparable date can be assigned to the limit at the head of the Humber Arm. The date, therefore, may be taken as that at which these fjord arms were ice-free. The reconstruction of ice-marginal positions during the retreat of the Newfoundland glacier from its maximum extent, discussed in Chapters Four and Eight, shows that ice margin probably lay across the mouth of the fjords at the time the St. George's Bay coastline was deglaciated, at 13,500-13,700 years B.P. By this reasoning, approximately 1,000 years were available for the cutting of the rock basins into the floors of the glacial valleys of the Bay of Islands. While this estimate is somewhat speculative, it cannot be extended very much. To do this would mean assuming that the outer margins of fjord glaciers were floated earlier than 13,500 years B.P. It is very unlikely that this occurred before that time since ice flow would have been vigorous, both through the valleys themselves and off the adjacent highlands.

The long profile of Humber Arm, Bay of Islands, is interesting with respect to Crary's mechanism of fjord formation. Humber Arm is twenty-eight miles long from the threshold to the mouth of the Humber

River. One major basin, 800 feet deep, is seen in the profile, extending for eight miles to the east of the threshold. Three smaller basins are eroded into this major basin. However, the floor of the arm is almost flat at a depth of 300 feet for 20 miles east of the basin. The absence of basins there may be explained in two ways. First, glacier ice in Humber Arm wasted rapidly into the low plateau areas bounding it, allowing rapid ingress of the sea the whole length of the valley. Second, in the upper 20 miles of the arm, the glacier tongue was instantaneously floated by penetrating sea water and basal erosion was halted. The similar elevation of the highest raised marine feature in Humber Arm and in Middle Arm, in which basins occur further toward the head, show that the sea first occupied the two arms at the same time. It is, therefore, easier to envisage a rapid disappearance of glacial ice from the upper 20 miles of Humber Arm, allowing the sea to reach its head.

The strikingly deep and steep-sided glacial troughs of Trout River Ponds (figure seven) and Serpentine Lake may also be fjords, although they are not sounded. The western part of Trout River Ponds lies close enough to the sea at present for it to have been flooded by late-glacial marine waters in contact with a glacier tongue occupying the trough. Rock basin formation could have occurred at the western end of the trough immediately prior to marine overlap. At the onset of marine overlap, the ice tongue in Trout River canyon seems to have become inactive since proglacial delta building into the sea began when the sea level had fallen to 115 feet from a marine limit at 195 feet.

The seaward end of Serpentine Lake is farther from the sea than Trout River Ponds, yet its surface is quite close to sea level. An ice tongue occupied the trough in late-glacial times and it deposited an end moraine between the present lake and the sea. If, prior to marine overlap, which took place to at least 130 feet above sea level, the sea was washing an ice-front just off the present coastline, it is hard to imagine rock basin formation occurring on the site of the present lake, 12 miles inland of the present coastline.

3. GLACIAL MODIFICATION OF DRAINAGE PATTERNS

Embleton and King (1968) mention the following causes of glacial modification of stream courses: ice damming, morainic obstruction, glacial erosion, and glacio-isostatic warping. To these can be added the cutting of new valleys (or the deepening of pre-existing ones) by ice-marginal and sub-glacial meltwaters, and the superposition of post-glacial stream courses on a cover of glacial deposits. Examples of all these causes are found in western Newfoundland. Some examples of drainage modification in the area show that more than one agency has been responsible. It is important to emphasize the varying scales of the modifications. These range from a radical alteration of major stream courses and changes in watershed area to minor deflections of low-order streams.

A small-scale topographic map of western Newfoundland (Plate I) reveals the general disposition of plateaus and lowlands whose position and form reflect major geological units. In pre-Quaternary times southwestern Newfoundland underwent a period of subaerial fluvial denudation, possibly beginning in the early Cretaceous, which was characterized by non-diastrophic periods of surface planation that were regularly interrupted by rejuvenating negative movements of the regional base level (Brookes, 1964). Naturally, the surface bevels slope in the direction of pre-Quaternary stream courses. These surface slopes, therefore, provide clues to the directions followed by pre-glacial drainage lines. Other evidence of former stream courses is suggested where present-day courses cross structural lineaments or dissect former divides.

At the highest level of generalization, divides between major drainage systems follow the crests of the plateaus, along the Long Range and Cape Anguille Mountains and the Bay of Islands massifs. Major structural troughs are occupied by high-order rivers such as the Codroy, Harrys, Serpentine, Trout rivers, and by segments of the Humber system and those rivers draining into Bonne Bay (Plate I).

The major pre-glacial drainage divide in western Newfoundland lay along the crest of the Long Range Mountains (Plate VII). One segment of it trended southwest to northeast from near Port aux Basques to east of the Bay of Islands, where a lower, north-trending segment linked it to another major segment along the crest of the Great Northern Peninsula. From the first segment of this divide, now represented by a line of residual summits at 1,900-2,200 feet, rivers drained into the Codroy River, St. George's Bay, Harrys River, and Bay of Islands lowlands in a general westerly direction. Planation surfaces atop the Long Range plateaus slope in this direction from those residual summits. To the east, rivers drained southeast to the southern shore of the island, east to join the Exploits River system, and northeast to the Atlantic Ocean down the major structural lowlands of Grand Lake and Deer Lake. Above the southeast side of Grand Lake a wide plateau at 2,000-2,300 feet shed drainage southeastwards to the Exploits system. From another branch of the major divide, west of Grand

Lake, streams drained north into the Bay of Islands, and west through the Serpentine River trench. From the segment of the major divide joining this part to that in the Great Northern Peninsula, streams flowed northwest through the Trout River and Bonne Bay troughs and southeast to join Atlantic drainage in the Deer Lake depression. In the Great Northern Peninsula drainage was from a divide closer to the western mountain front than the present one, to the Gulf of St. Lawrence on the western side, and to the Atlantic on the eastern side.

Major breaches of these divides have occurred as a result of glaciation. The most radically affected area is that node from which drainage once flowed in all directions, namely the Grand Lake-Lower Humber area. It will be well attested in Chapter Four that at the maximum of glaciation ice moved westwards over the west coast of the island from an ice-shed along the Long Range Mountains. Valleys indented into the height of land along those mountains would certainly be vulnerable to overdeepening by ice flowing strongly to each side of it. The erosive activity probably was more intense on the western side of the divide because of a steeper ice-surface profile that increased ice flow rates. A steeper profile would have been the result of a steeper land-surface slope to the Gulf of St. Lawrence and, later, of the rapid calving of the ice front into the late-glacial sea waters. Valley glaciers with steep profiles endured in the troughs dissecting the Long Range and Bay of Islands Mountains while the sea was overlapping the present Gulf shoreline.

The upper Grand Lake breach of the divide was the result of erosion by a glacier tongue which cut a U-shaped trough in an east-west direction from the icefields atop the Long Range into the Harrys River lowland (figure 16). A stream valley which fed waters westwards into that lowland was overdeepened, and the valleys each side of the interfluve which now forms Glover Island in Grand Lake were strongly overdeepened. Upon wastage of the glacier tongues in these troughs, and the blockage of the western outlet of Grand Lake by ice-contact glacial debris, drainage to the southwest was beheaded. The northeasterly segment of a stream which had once continued southwestwards into St. George's Bay now flows into the western end of Grand Lake. That sub-glacial drainage was directed to the west is well shown by the trend of an esker system leading from the lake to the west, and then southwest, into the Harrys River lowland.

Before discussing the Grand Lake drainage further, the Humber River breach must be considered as an example of a similar development. East of the head of the Bay of Islands, the Long Range Mountains trend southwest to northeast, with summits at 1,800-2,000 feet. From this divide, pre-glacial drainage flowed west into the Bay and northeast into the Deer Lake lowland and thence to the Atlantic. Breaching of the divide by a glacier tongue produced a through-valley in which water from the northeast, rising only eight miles from the Atlantic shore, now courses southwest through the glacial gorge into Humber Arm, Bay of Islands (figure 17).

The Grand Lake and Humber River drainage basins are interconnected and find their present outlet in the Bay of Islands assisted by another of the causes of drainage modification during glaciation - isostatic warping. The topographic map (Plate I) shows that the logical outlet for drainage east of the Long Range, from Grand Lake and the Humber River, would be into White Bay. In fact, around the head of that bay, streams rise as close as

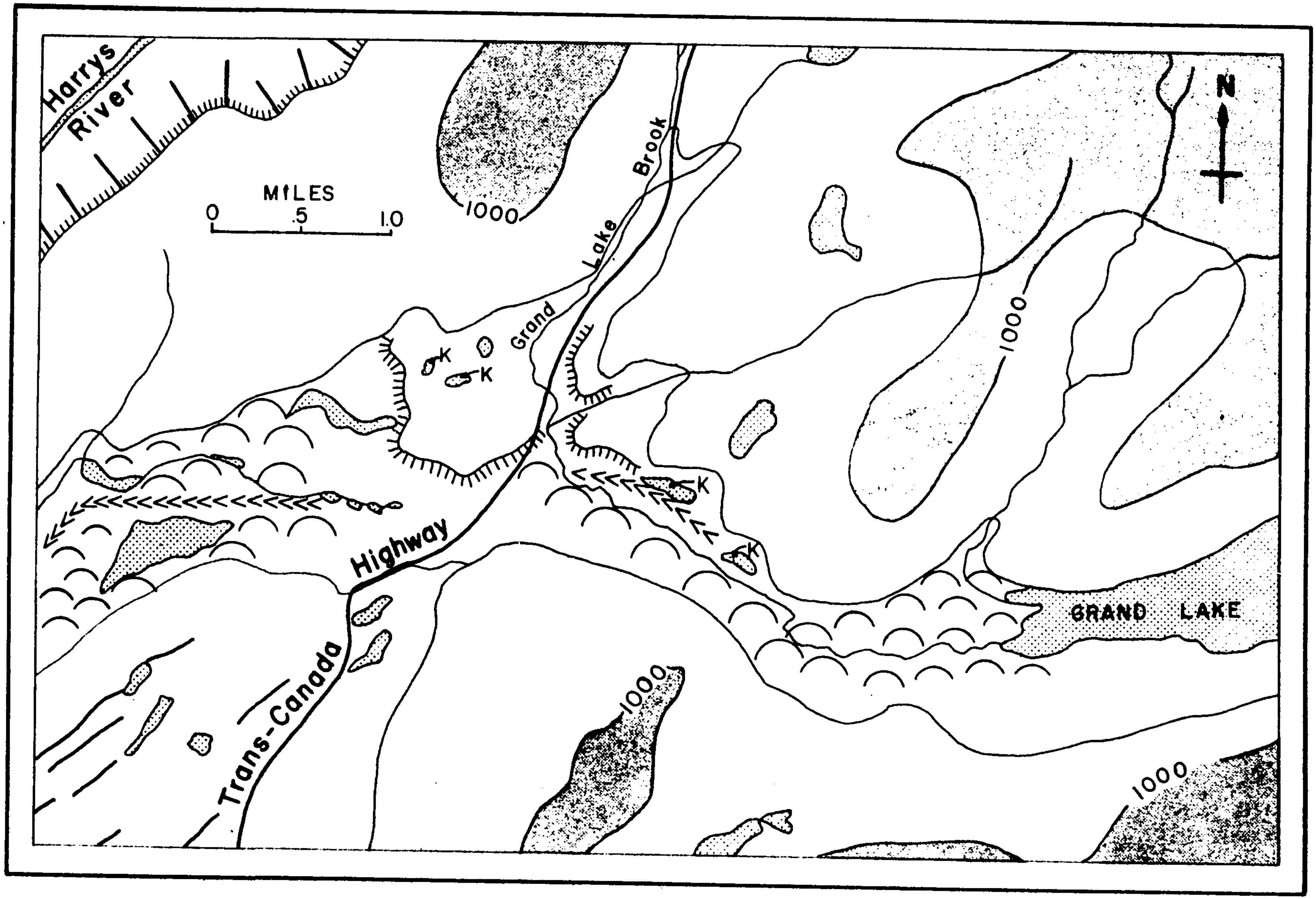


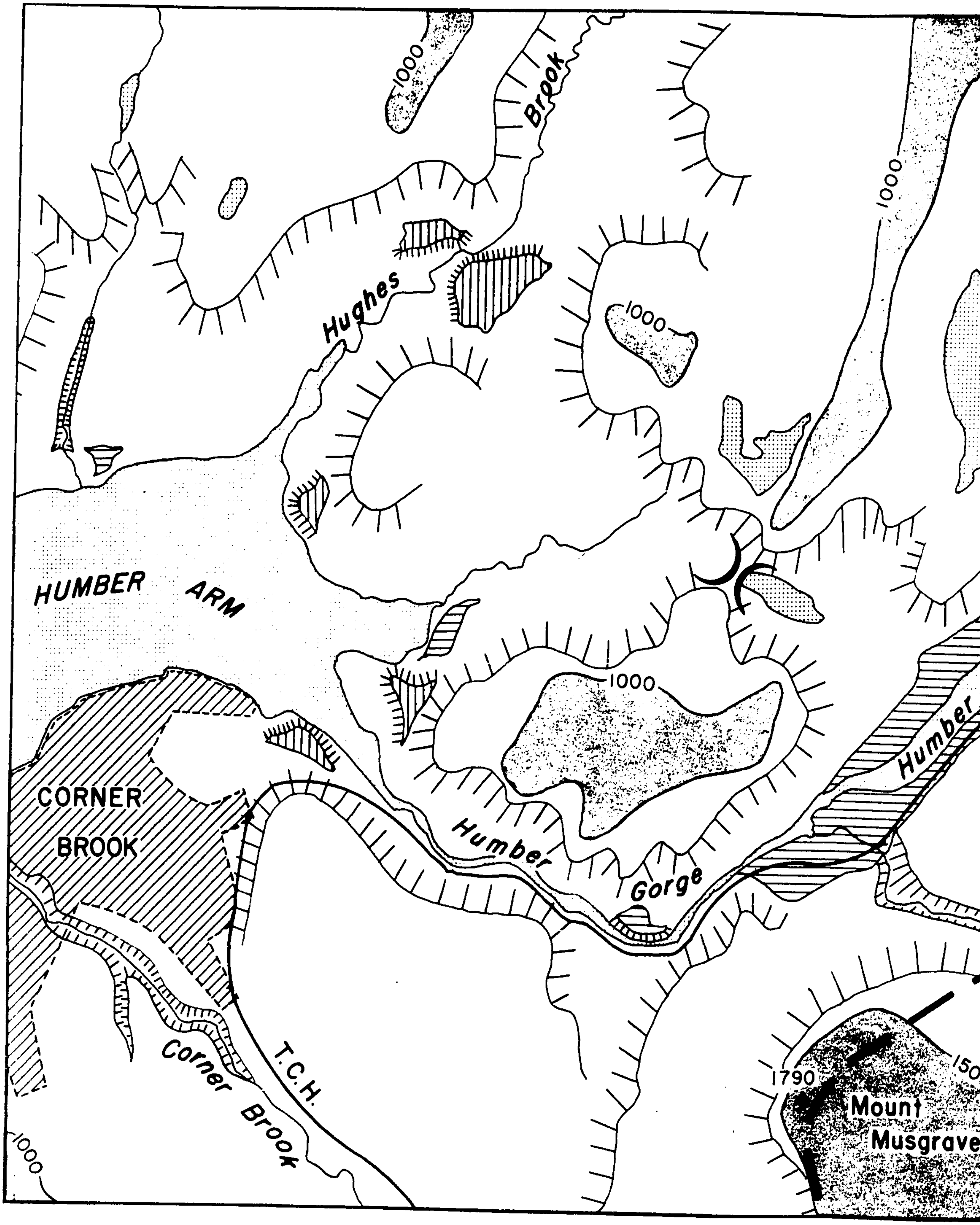
Figure 16. THE GRAND LAKE GLACIAL BREACH

eight miles from the Atlantic Ocean and flow 60 miles away from it to the Gulf of St. Lawrence. Raised marine features around the Bay of Islands are near 160 feet at their highest whereas, around White Bay and Green Bay on either side of the Burlington Peninsula, they reach 250 feet (MacClintock and Twenhofel, 1940; Lundquist, 1965). This 90-foot difference in recorded isostatic rebound is large in comparison with the 200-foot total fall of the Humber River from its middle reach, northeast of Deer Lake, to the sea. The real difference in isostatic uplift between White Bay and the Bay of Islands is more than 90 feet, since the former area was deglaciated 1,000 years later and has, therefore, recorded less uplift than the Bay of Islands.

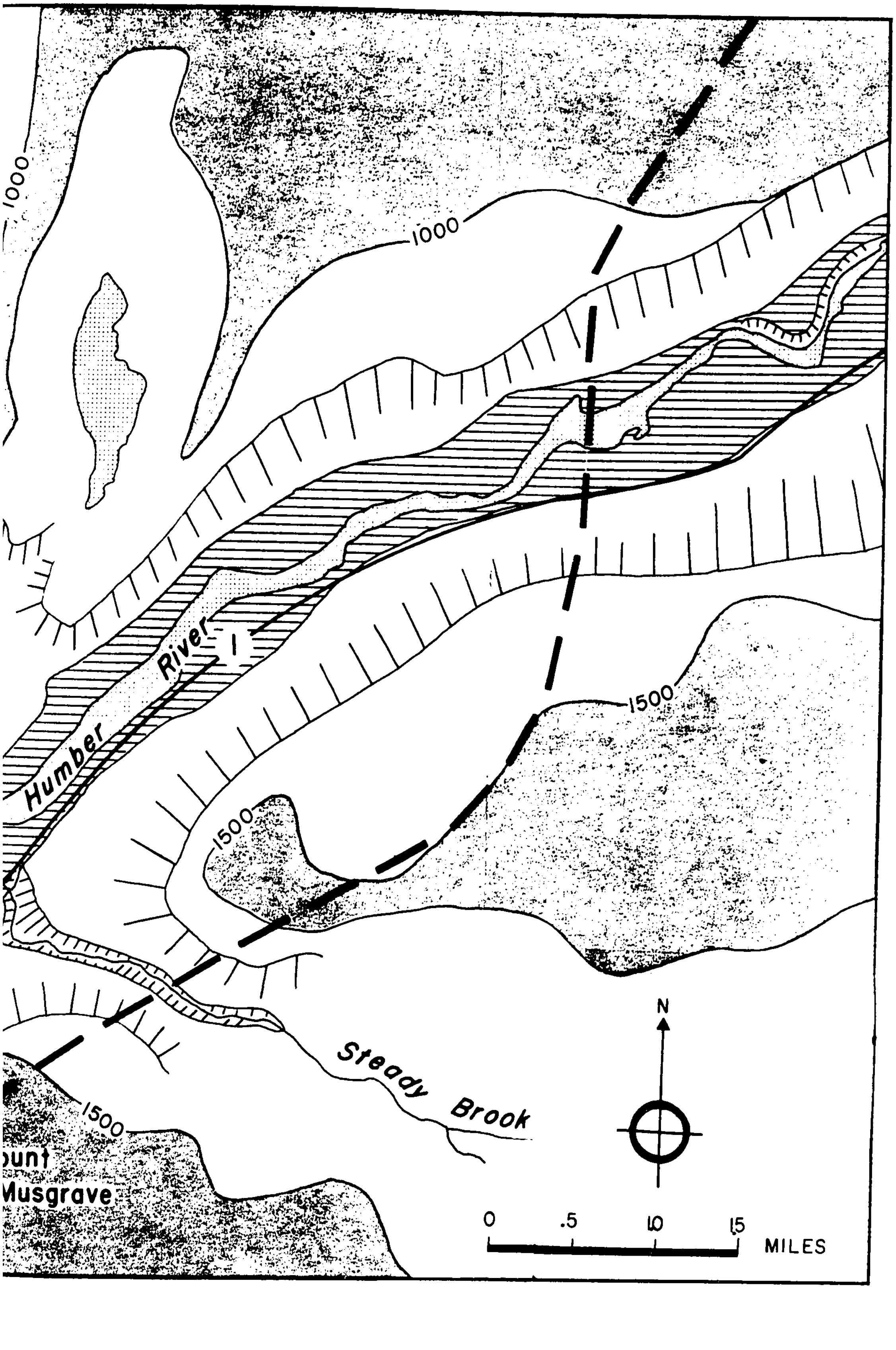
The same factor has contributed the waters of Sandy Lake, Sheffield Lake and Birchy Lake to lower Grand Lake and has caused the latter to spill over an ill-defined watershed into the Humber at Reidville, five miles upstream on the Humber from Deer Lake. The two glacial breaches, assisted by isostatic tilting generally towards the north, have produced the major drainage changes in the western part of the island and have enlarged the present Humber River drainage to approximately eight times its pre-modification area.

All other glacial modifications of drainage are minor by comparison with those discussed above. Perhaps the least minor, and the most problematical due to a lack of evidence, is the case of the drainage across the lowland between the Long Range and the east shore of St. George's Bay. Any direct evidence of the courses of pre-glacial streams across the lowland has been obliterated by glacial deposits. One or two of these rivers show angulate directional changes as they cross the fault-line bounding the

Figure 17. THE LOWER HUMBER RIVE



ER RIVER GLACIAL BREACH



mountains. Otherwise all these rivers follow sub-parallel courses with no hint of control by bedrock structures beneath the glacial deposits. These structures have been mapped as predominantly southwest-to-northeast-trending folds (Riley, 1962), with dips from 10 to 85 degrees and normal faults. It would be logical to expect that pre-glacial drainage would show some adjustment to these structures, with some sections oriented parallel to the present shoreline. The present pattern, whether or not it differs from the pre-glacial one, is due to superimposition on a seaward-sloping cover of morainic and ice-contact debris up to 150 feet thick.

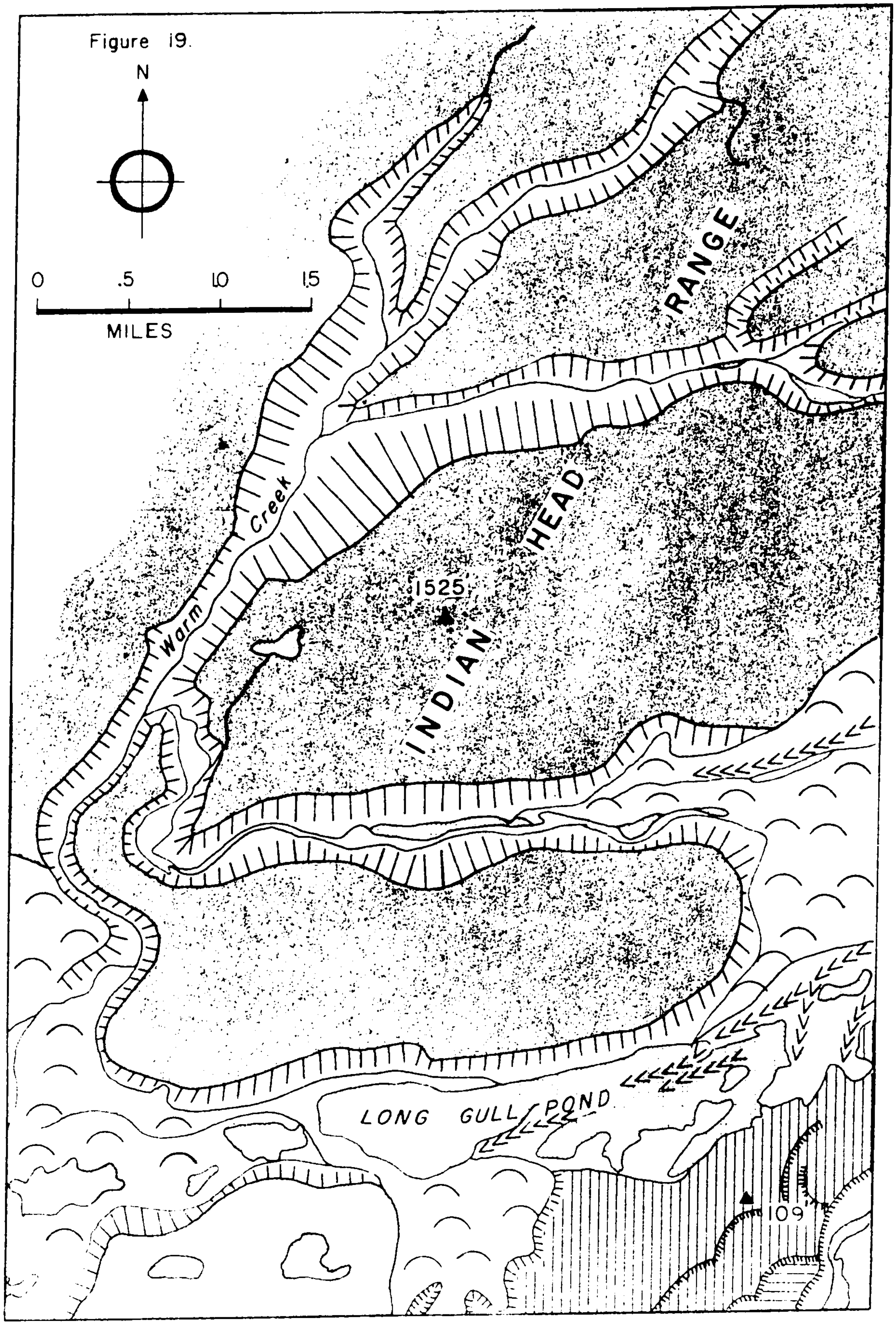
The southern part of this lowland is cut off from the Codroy lowland by an upland remnant of the Cape Anguille Mountains, composed of Mississippian sandstones. Planation surfaces at 1,400-1,700 feet atop the mountains and this remnant show that drainage once flowed both to the northwest into St. George's Bay, and southwest into the Codroy system. A stream from the Long Range plateau also drained northwest through a valley separating the two plateaus, but this has been obstructed by glacial debris and drainage has turned southwestwards through a right angle to join the Codroy drainage (figure 18). The defile now occupied by Codroy Pond is a glacially and glacio-fluvially overdeepened descendant of a saddle from which streams drained northeast and southwest. The valley is now plugged with glacial debris, and underfit streams flow in opposite directions to join Highlands River and the Codroy River.

Although glacial meltwaters probably contributed to some extent to watershed breaching by ice, a case exists in western Newfoundland in which it was solely responsible for a small modification of a drainage pattern. Across the southern and middle sections of the Indian Head Range,

Figure 18. **THE CODROY RIV**

ROY RIVER GLACIAL BREACH

at the head of St. George's Bay, three valleys cut across resistant igneous rocks to depths of 700 feet below the upland surface (figure 19). The southernmost valley occupied by Long Gull Pond, is a glacially eroded trough, but the northern pair are solely the product of glacial meltwaters. Of these, the southern one lies at 700 feet below the general surface. It slopes evenly through 180 feet in six miles from east to west. An esker which leads into the valley and glacial debris which partially blocks a section of it, indicate a sub-glacial origin. It also has a remarkable incised meander loop cut 300 feet through bedrock at its western end, indicating powerful water erosion by a debris-charged sub-glacial stream under hydrostatic pressure. The northern channel is less impressive. The average slope is 80 feet per mile, with an upper section sloping at 50 feet per mile and a lower at 100 feet per mile. It has no esker leading into it from the east since it does not open to the Harrys River lowland. Only its subdued similarity to the southern channel and its steep, V-shaped cross-profile link the two genetically. These valleys, and that glacial trough now occupied by Long Gull Pond, fed debris-charged meltwaters from the ice in the Harrys River lowland into the sea to build the large delta near Harmon Field. The two southernmost valleys presently lead drainage from the east across the pre-glacial divide which was oriented south-to-north over the Indian Head Range.



The Indian Head Range watershed breaches

PART TWO

Introduction

Part Two deals with the chronologic aspects of the study. Chapters Four, Five and Six treat each of the three glacial events recorded in the sediments and landforms of the study area. MacClintock and Twenhofel (1940) gave the name "stage" to each of these three glacial events. In order to avoid the implication of a definite rank within the stratigraphic scheme erected in the North American continental interior, the term "event" is used in this study. The names which MacClintock and Twenhofel gave to their "stages" are here retained for the "events" for the sake of continuity of nomenclature. Chapter Seven deals comprehensively with the postglacial marine phase and isostatic uplift. In that chapter, some of the evidence used in the previous three chapters to arrive at a chronology of glacial events is used again to view the phenomena associated with marine overlap as a whole. Chapter Eight summarizes the main conclusions of the study and refers to the results of recent studies of glaciation in the mainland Atlantic Provinces of Canada. Some problems for future research are mentioned in that discussion.

CHAPTER FOUR

The St. George's River Drift Event

1. THE EXTENT AND CHARACTER OF THE LANDFORMS AND DEPOSITS

The name "St. George's River Drift" was given by MacClintock and Twenhofel (1940) to the lowest Quaternary unit exposed in coastal cliffs and inlets bordering St. George's Bay. The place of this unit in the scheme erected by those workers has been discussed in Chapter Two. The name of the drift is taken from the inlet which extends tidewater seven miles to the east of Stephenville Crossing, near the head of St. George's Bay. On each side of this inlet the drift underlies a wide area beneath a cover of blanket bog. Elsewhere in southwestern Newfoundland, surface exposures of the St. George's River Drift are all in till which lightly mantles the margins of bedrock plateaus and valley sides, particularly around the Port au Port Peninsula and along the sides of the fjords further north. In these situations, late-glacial marine waters have modified the surface character of the till, washing out the fines and exposing large boulders, and etching small bluffs into the till at intervals from the marine limit to present sea level. The upstanding plateaus of the area are mostly bare of glacial deposits. On extensive plateaus of low relief in the Long Range Mountains are seen patches of ice-stagnation debris which relate to the latest phase of glaciation, and are not correlative with the St. George's River Drift. The plateaus of the Bay of Islands Mountains and the Port au Port Peninsula are practically devoid of glacial deposits. Actively flowing ice crossed these, however, and, if this did not succeed in inhibiting subglacial deposition, any deposits have been subsequently removed by postglacial fluvial erosion.

It is in the coastal cliffs around St. George's Bay that the St. George's River Drift is best exposed to view. As the lowest of the glacial units, it occupies a low position in the cliffs, except where Carboniferous or Ordovician bedrock rises higher. Generally, the bedrock surface lies close to sea level. Thus, except where it falls below sea level itself, the drift is the easiest of the units to examine (figure 20).

Deposits comprising this drift are of two types. The commonest is a compact, massive till, rich in clay and silt, which appears reddish-brown, or greyish-brown when wet. The massive structure gives it sufficient strength to stand in near-vertical faces when freshly attacked by high seas, and it therefore is visibly prominent at the foot of the cliffs by comparison with the looser structured sands and clay-silts above. Less common are the pockets of loosely structured, coarse, ice-contact gravels which occur as hummocks on the till surface in a few places along the St. George's Bay shore.

Textural analyses of five till samples from widely separated localities attest to the homogeneity of the till (figure 21)[1]. The relative abundance of finer particles in this till compared to those in the younger Robinsons Head Drift till (figure 22) is related to two facets of its genesis. Firstly, it is a lodgement till, consisting of a mix of particles of all sizes, with no selective removal by englacial waters.

---

1. The textural analyses were done on the fraction passing the 3/8-inch sieve, using the method outlined in Dawson (1959).

Figure 20: Compact lodgement till of St. George's River Drift in base of cliff near mouth of Fishels River. Pocket of ice-contact stratified debris at base.

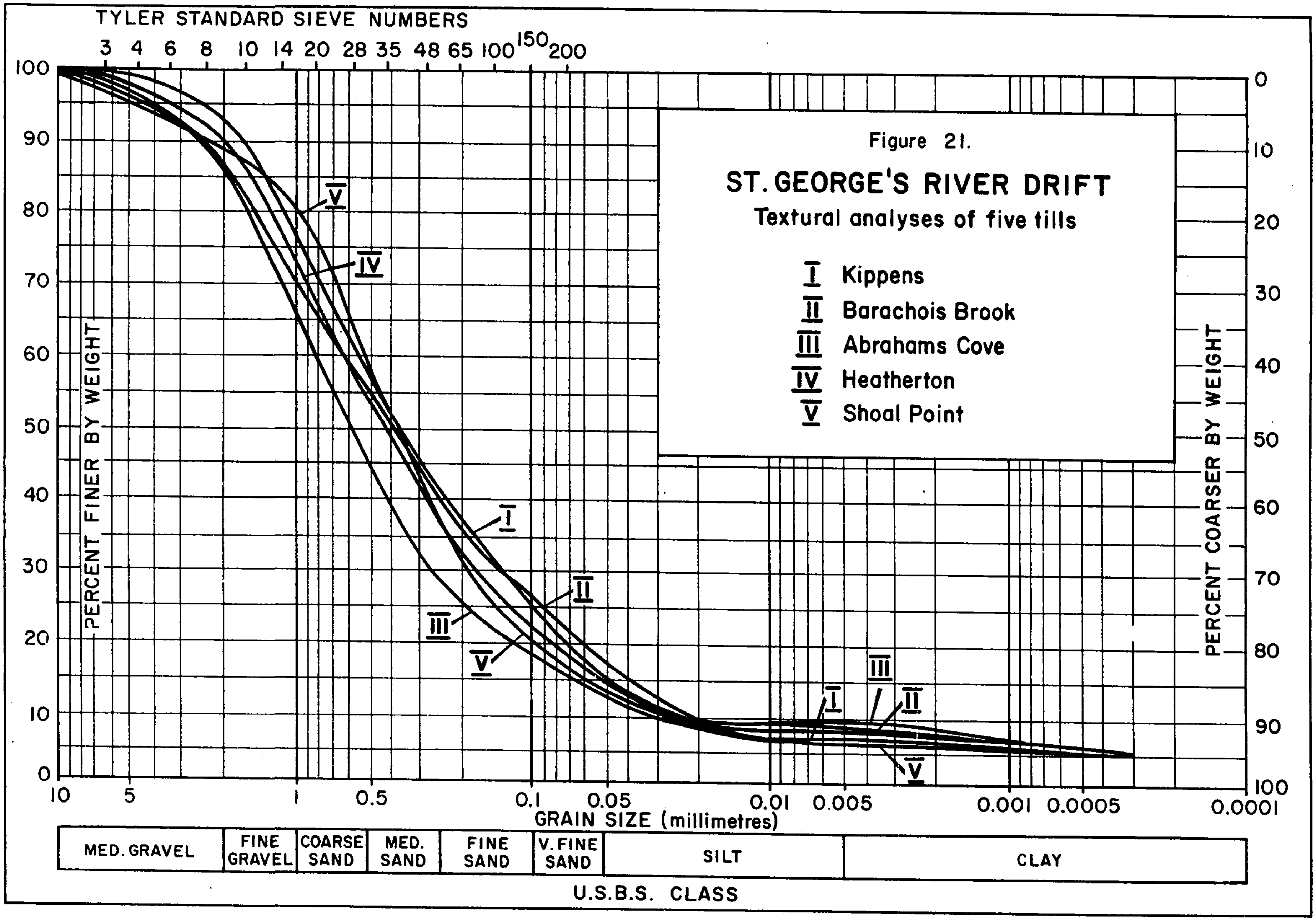
Figure 20: Compact lodgement till of St. George's River Drift in base of cliff near mouth of Fishels River. Pocket of ice-contact stratified debris at base.

Secondly, ice that deposited this till, as will be shown in this chapter, passed from east to west across Carboniferous clastic sedimentary rocks in which shales are common. Larger clasts in the till are predominantly of igneous and metamorphic rocks, the source outcrops of which occur in the mountains bordering the St. George's Bay lowland to the east and north. Many of these clasts can be traced to bedrock source areas to provide evidence of ice-flow directions during till deposition (see section 2). The equidimensional form of the clasts made it impossible in the time available to obtain a sufficient number of them suitable for till macro-fabric analysis. It is generally recognized that fifty pebbles is the minimum number required from a site to perform the analysis, but far fewer than that were collected from all five sites at which the samples were obtained for textural analysis. The plastic consistency of a basal debris-water-ice mixture, such as has been postulated to have caused the smoothing and plastic scouring of bedrock surfaces in the Port au Port area, would certainly have oriented any elongate, flattened pebbles parallel to the ice-flow direction, but, unfortunately, such pebbles are scarce.

The pockets of ice-contact gravels, seen in such localities as near Romaines Brook and Robinsons Head, lack the silt and clay grades of the underlying till. Their hummocky form indicates an origin as kames deposited between stagnant ice blocks. As the larger clasts are of the same rocks as in the till, a similar source area is indicated.

2. THE PROVENANCE OF THE DEPOSITS AND DIRECTIONS OF ICE FLOW PRIOR TO MARINE OVERLAP

Clues to the provenance of the St. George's River Drift have been


TYLER STANDARD SIEVE NUMBERS
3 4 6 8 10 14 20 28 35 48 65 100 150 200
PERCENT FINER BY WEIGHT
PERCENT COARSER BY WEIGHT
GRAIN SIZE (millimetres)
10 5 1 0.5 0.1 0.05 0.01 0.005 0.001 0.0005 0.0001
MED. GRAVEL
FINE GRAVEL
COARSE SAND
MED. SAND
FINE SAND
V. FINE SAND
SILT
CLAY
U.S.B.S. CLASS
Figure 21.
ST. GEORGE'S RIVER DRIFT
Textural analyses of five tills
I Kippens
II Barachois Brook
III Abrahams Cove
IV Heatherton
V Shoal Point

gained from tracing erratic till boulders to their source outcrops. This evidence indicates ice-flow directions that are corroborated by the directions of glacial striae. The latter also indicate the changes in directions of ice flow with time. Both lines of evidence and conclusions drawn from them are shown in figure 23, which should be referred to in the following paragraphs.

Large clasts in the St. George's River Drift till are often of such distinctive rock types that they can be traced to a unique source area, shown in Riley (1962). A boulder of serpentinized dunite, an olivine-rich rock, lies on a sloping till surface above the village of Mainland. Many smaller boulders of the same rock occur among other erratic boulders of granite and local limestone fragments atop the 1,200-1,300-foot summit of Table Mountain, Port au Port. The nearest source area for this rock is in the Lewis Hills to the northeast of Mainland and north of Port au Port. The Blow Me Down Mountains and mountains north of the Bay of Islands are also largely built of the same rock, but the direction of transport of the erratics does not change substantially if they are traced to those source areas. The distances of transport increase, but here conservative interpretations have been favoured.

On the southern shore of the Port au Port Peninsula, at Campbell's Cove, a large erratic boulder of anorthosite, six feet on a side, lies on the modern beach near the level of mean high tide. It is unlikely that sea-ice could have transported it, although it must be admitted that little is known of the thickness, strength, and movements

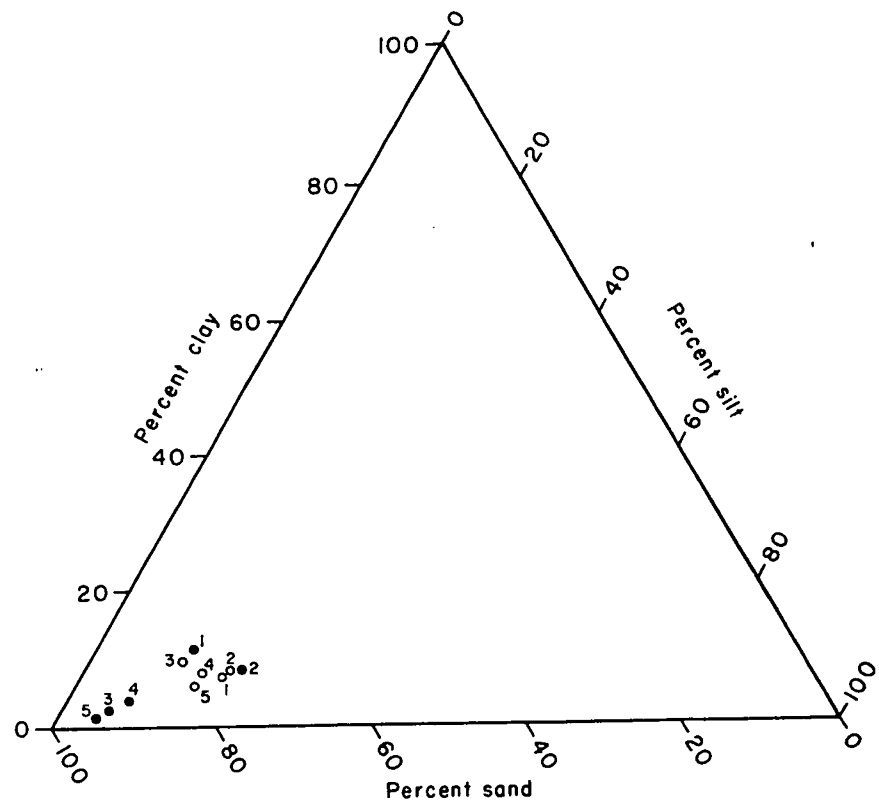


Figure 22.

TEXTURAL ANALYSIS OF TILLS
FROM ST. GEORGE'S BAY AREA

o St. George's River Drift tills

• Robinsons Head Drift tills

For identification see grain-size curves

U.S. Bureau of Soils Classification

of sea-ice in the area. The anorthosite is distinctive in its mauve colour and the inclusion of large crystals of titaniferous magnetite (Riley, 1962). The only surface outcrop of this rock in the study area lies astride the valley of Flat Bay Brook, east of St. George's. A transport of at least twenty-two miles from the east is indicated for the Campbell's Cove erratic.

Many smaller boulders of this anorthosite occur in the St. George's River Drift till, in the coastal cliffs and on the modern beach, due west of the outcrop of the rock, between Little Barachois Brook and just south of the mouth of Fishels River. Westward-flowing ice is again indicated. Another quite different anorthosite occurs at the southern end of the Indian Head Range, near Stephenville. It is white in colour, with large albite crystals and inclusions of enstatite pyroxene (Riley, 1962). Boulders of this anorthosite occur in the St. George's River Drift till west of Stephenville, and a few large boulders, up to six feet on a side rest on the modern beach near low-tide level (figure 24). The latter have occupied a low-tide position for the five years this study has been in progress, so that it is unlikely that sea-ice transported them there. They, therefore, have been washed from the till in the cliffs. A westward transport of six miles has occurred.

At the southern apex of the drift-covered St. George's Bay lowland, south of Highlands, the St. George's River Drift till exposed in coastal cliffs contains large, striated boulders of reddish-brown, arkosic sandstone belonging to the Mississippian Anguille Group. The nearest source outcrop of these rocks is in the Cape Anguille Mountains which rise from the lowland one to four miles inland of the coast here. A short

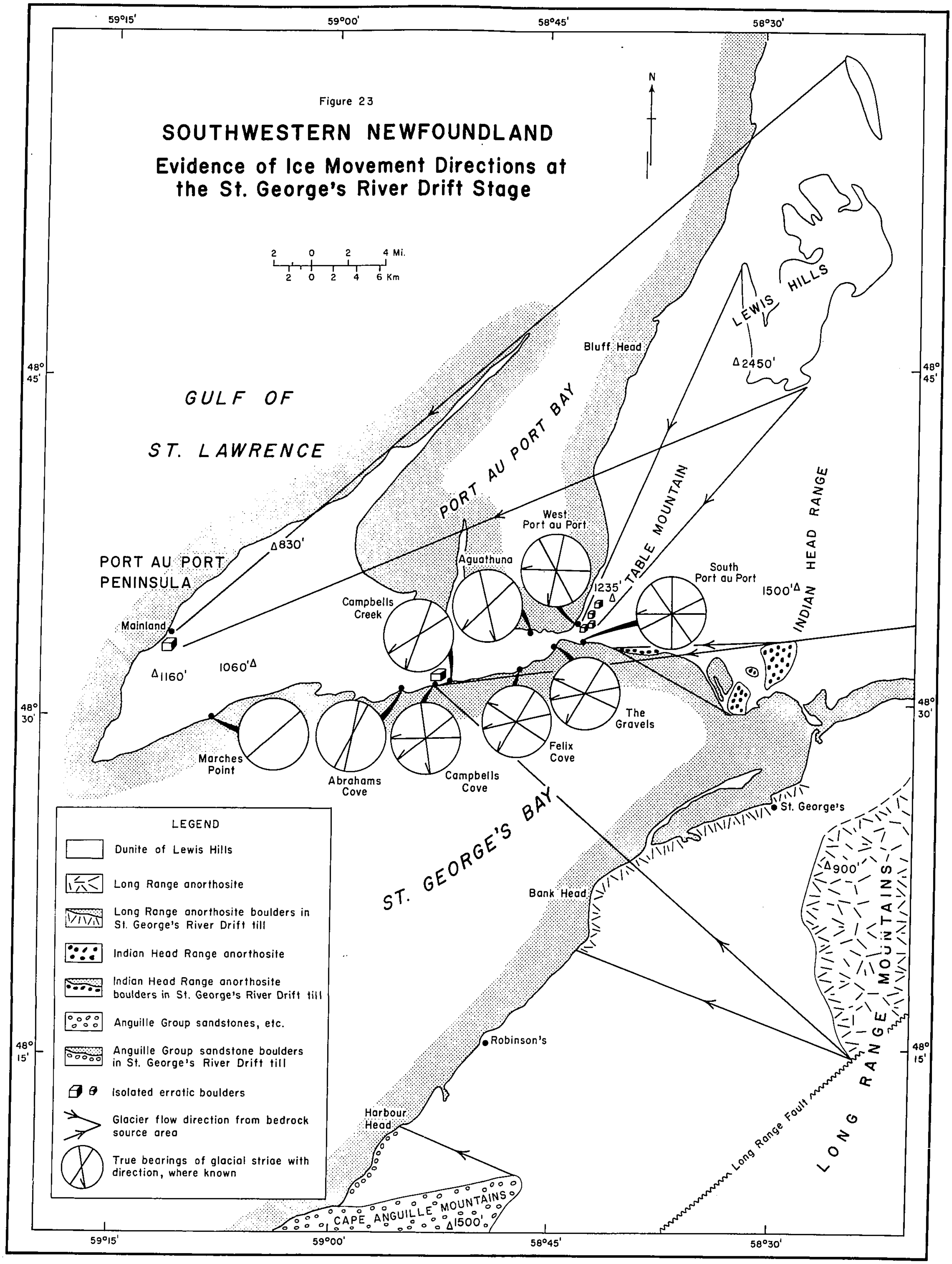


Figure 23
SOUTHWESTERN NEWFOUNDLAND
Evidence of Ice Movement Directions at the St. George's River Drift Stage
59°15'
59°00'
58°45'
58°30'
48° 45'
48° 30'
48° 15'
N
2 0 2 4 Mi.
2 0 2 4 6 Km
GULF OF ST. LAWRENCE
PORT AU PORT BAY
PORT AU PORT PENINSULA
LEWIS HILLS
2450'
Bluff Head
TABLE MOUNTAIN
INDIAN HEAD RANGE
West Port au Port
South Port au Port
1500'
1235'
830'
Aguathuna
Campbells Creek
Mainland
1160'
1060'
Marches Point
Abrahams Cove
Campbells Cove
Felix Cove
The Gravels
ST. GEORGE'S BAY
St. George's
900'
Bank Head
LONG RANGE MOUNTAINS
Robinson's
Harbour Head
Long Range Fault
CAPE ANGUILLE MOUNTAINS
1500'
LEGEND
Dunite of Lewis Hills
Long Range anorthosite
Long Range anorthosite boulders in St. George's River Drift till
Indian Head Range anorthosite
Indian Head Range anorthosite boulders in St. George's River Drift till
Anguille Group sandstones, etc.
Anguille Group sandstone boulders in St. George's River Drift till
Isolated erratic boulders
Glacier flow direction from bedrock source area
True bearings of glacial striae with direction, where known

distance of transport from the southeast is indicated.

The foregoing evidence of ice-flow directions is supplemented by that from glacial striae in the Port au Port area. The conditions under which these were formed have been discussed in Chapter Three. Glacially smoothed limestone ledges in the Port au Port area show many of the micro-features of glacial erosion. Since these ledges are overlain by till of the St. George's River Drift, the erosional features can be assigned to the glacial episode during which the latter was deposited. Arguments for a glacial origin have also been forwarded in Chapter Three. Figure 23 shows that striae are grouped into three sets, which may not all be present at a locality. The strongest set is that from predominantly east to west (figure 25). Weaker striae, with more variable directions, are grouped around directions slightly east and slightly west of north towards the south. Where the two north-south sets are seen together, they cross each other with such irregularity that they can be assigned to the same ice mass moving from the same outflow centre, but shifting direction slightly due to the vagaries of topography and glaciological conditions. When one or both of the north-south sets are seen with the east-west set at any locality, the former cross the finer east-west striae and the gentler facets of crescentic gouges, but do not always continue across the entire width of the deeper and wider east-west striae and grooves. This latter relation may have been that which prompted MacClintock and Twenhofel (1940) to assign north-south striae to an earlier age than east-west ones. However, not only do north-south striae cross the finer east-west ones, but the fact that they do not cross stronger east-west striae can be readily explained. I interpret the relation as due to a decrease of stress and

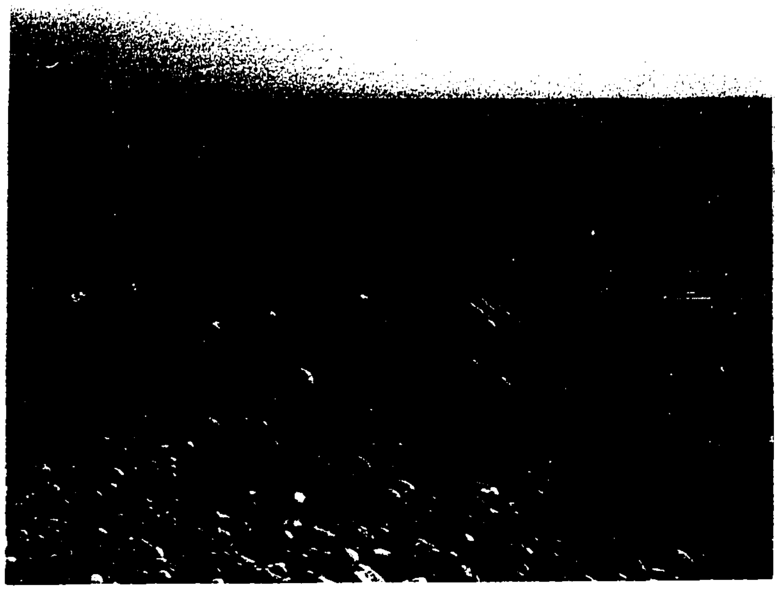


Figure 24: Indian Head Range anorthosite erratics washed from St. George's River Drift till on beach west of Stephenville.

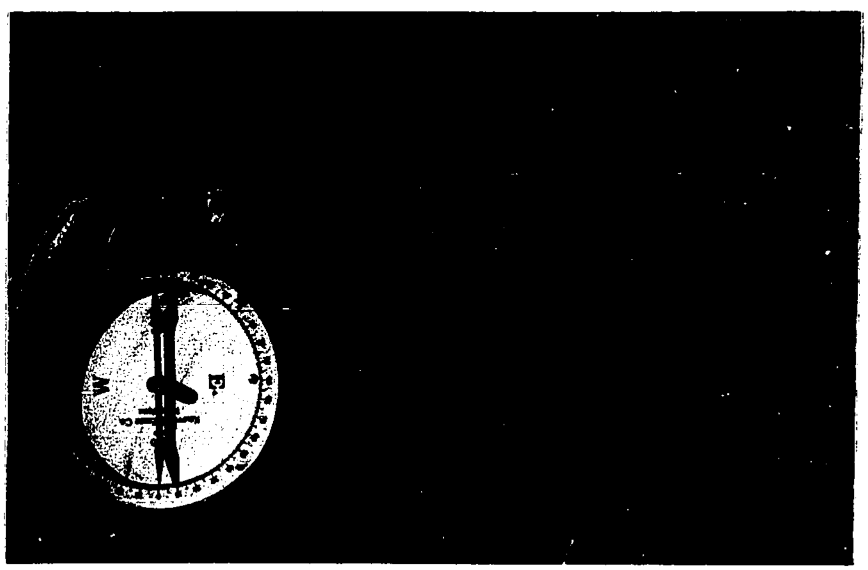


Figure 25: Crossing striae at Campbells Creek, Port au Port Peninsula, indicate ice flow first from 050°, and later from 000°True.

Figure 24: Indian Head Range anorthosite erratics washed from St. George's River Drift till on beach west of Stephenville.

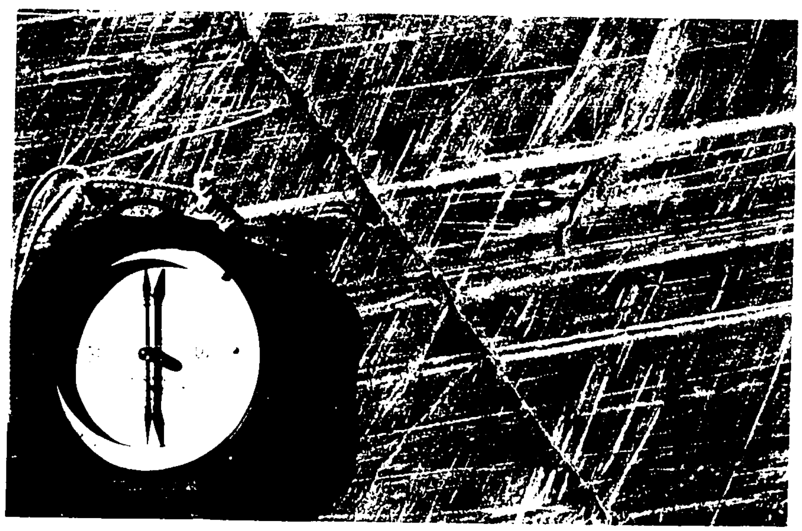


Figure 25: Crossing striae at Campbells Creek, Port au Port Peninsula, indicate ice flow first from 050^{o}, and later from 000^{o}True.

striating ability at the base of ice moving southwards when it reached the lip of a deep east-west striation. Passive ice passed across it and stress was built up again on the opposing edge.

Erratic boulder provenances, together with striae directions and ages indicate that strong ice-flow from a centre over the Long Range Mountains moved westwards into the Gulf of St. Lawrence at the Wisconsin maximum. At a later time, immediately prior to marine overlap, more variable ice-flow directions prevailed, with flow generally north-to-south over the Port au Port Peninsula, but continuing towards the west over the east shore of St. George's Bay. Figure 23 shows the coincidence of many striae directions with directions of erratic till boulder transport.

The foregoing conclusion, that southwestern Newfoundland was glaciated by an ice cap centred over the Long Range Mountains at the Wisconsin maximum, can be extended to most of the remaining area of the island. Some published evidence exists in the northern part of the Great Northern Peninsula that Labrador ice invaded the island there (Cooper, 1937; Prest, et al., 1968), and preliminary results of investigations by the Geological Survey of Canada also suggest that Labrador ice affected the northern part of the peninsula (Grant, 1970b). However, south of Bonne Bay, there is no evidence of the former presence of Labrador ice, so it is logical to assume that the ice cap over the Long Range Mountains, south of the northern peninsula, was responsible for the glaciation of the eastern and southern part of the island. With such a strong influence over ice flow directions in southwestern Newfoundland, it is difficult to imagine that, over the Long Range of the Great Northern Peninsula, a local ice cap would be dominated by northwest to southeast flow of Labrador ice.

This ice would have had to cross the 1,000-1,500-foot fault-line scarp bordering those mountains and the deep channel of White Bay if it was to affect eastern Newfoundland. Further, there is no evidence east of the Long Range of northwest to southeast ice flow which cannot be attributed to flow from an ice cap situated over the western highlands or immediately to the east of them.

The view that the island of Newfoundland supported its own ice cap at the Wisconsin maximum, and that evidence for the invasion of Labrador-derived ice is restricted to the northern part of the Great Northern Peninsula, has been the most prevalent one since the late nineteenth century. In Chapter Two it was seen that some of the earlier evidence upon which this view was based cannot be seen as convincing today. Since 1940, however, there has been less certainty about the relative influence of Labrador ice on the glaciation of Newfoundland. The conclusions of Flint (1940) were in direct opposition to those recorded earlier, in that they called for an invasion of the western part of the island by Labrador ice to account for the alleged trend of isostatic warping. The quotation from Flint (1940) in Chapter Two places a dominating influence on Labrador ice over the west coast of the island at the Wisconsin maximum.

The detailed studies of MacClintock and Twenhofel (1940) provided evidence of radial flow from an island ice cap prior to marine overlap. However, the north-south striae in the Port au Port area provided an opportunity for them to harmonize this idea with that of overriding or, at least impinging, Labrador ice favoured by Flint. They interpreted the north-south striae at Port au Port as older than the east-west ones and called upon Labrador ice, or island ice deflected southwards by it, to cross the Port au Port area at the Wisconsin maximum, and upon later radial flow to produce east-west striae at Port au Port prior to marine overlap.

My re-interpretation of the discrete ages of glacial striae in the Port au Port area, and the evidence gained in this study of erratic till boulder provenances, discount the former presence of Labrador ice, at least on present-day land areas. The question is now raised as to whether this ice lay offshore and was active enough to deflect island ice flowing west off the western highlands of Newfoundland to the south across the Port au Port Peninsula. Firstly, the southwestward movement of erratic boulders, shown in figure 23, indicates that the south-flowing ice was island ice and not from Labrador. Secondly, since this movement was later and weaker than the strong westward flow at the Wisconsin maximum, it is unreasonable to suppose that Labrador ice deflected Newfoundland ice southwards. If this occurred, it would be more likely to have happened at the earlier and stronger phase when both ice sheets would have been more active.

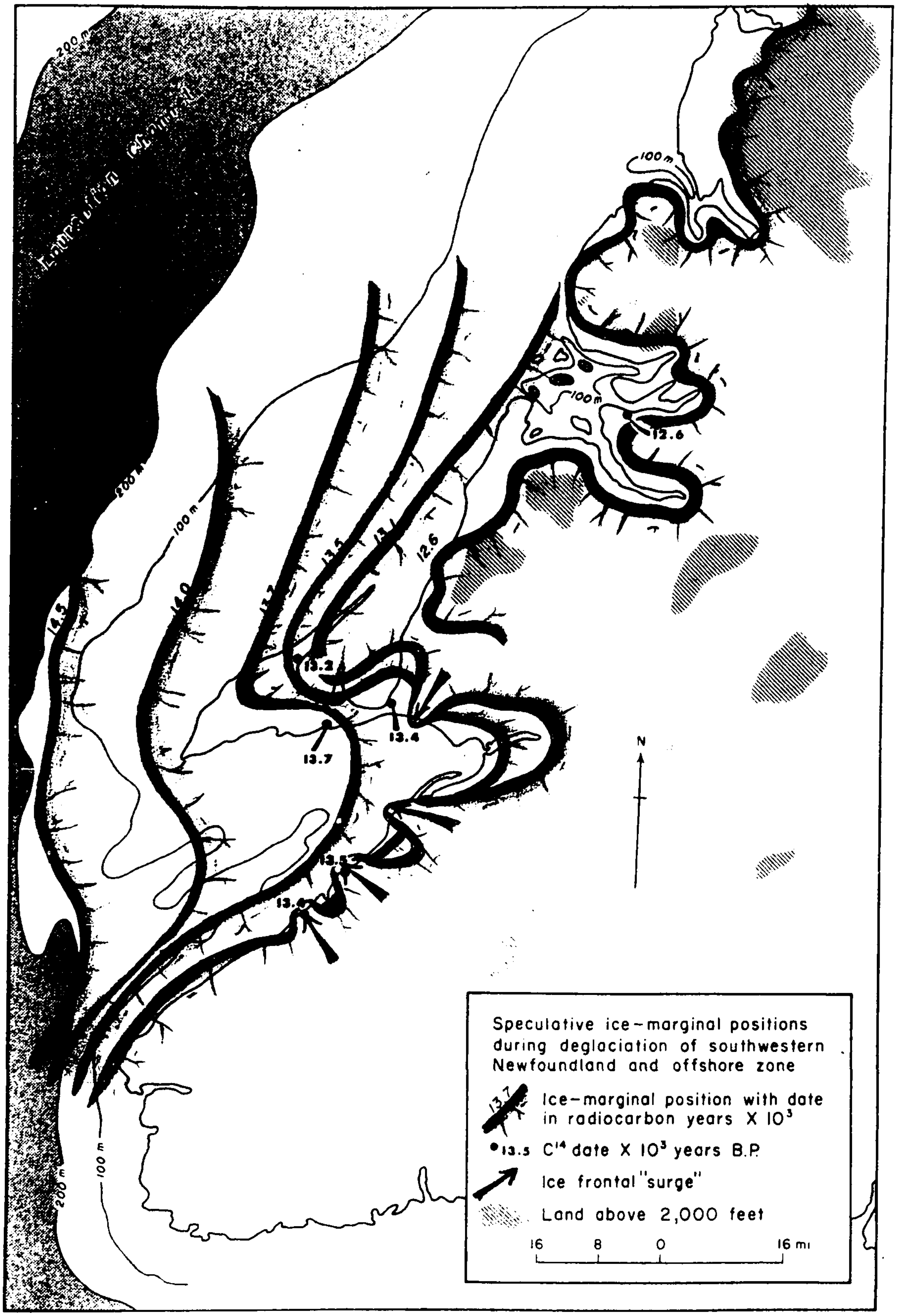
A southward movement of Newfoundland ice over the Port au Port Peninsula can be accounted for by a combination of two factors. Firstly, a decline in the activity of ice from the main island outflow centre over the Long Range Mountains would have isolated subsidiary centres over the Bay of Islands Mountains - for example, in the Lewis Hills (Plate I). These stand 300-500 feet higher than the 2,000-foot summits of the Long Range, and at present receive comparable amounts of snowfall. While the Long Range receives a maximum of over 200 inches, the Bay of Islands Mountains receive 150-175 inches (Hare, 1952). Ice flow from the subsidiary centres would have been radial, with a strong component to the south, due to the changing orientation of the ice margin in St. George's Bay.

This latter phenomena is a second important consideration in explaining ice-flow directions over the St. George's Bay shoreline immediately prior to marine overlap. The map of Prest (1969), "Retreat of Wisconsin and Recent Ice in North America," shows a tongue of ice in Cabot Strait at 14,500-15,000 years B.P., the flow of which was guided by the configuration of the Laurentian Channel. By 13,500 years B.P., marine waters flooding the Gulf of St. Lawrence were washing the newly deglaciated shores of St. George's Bay. Evidence for this will be presented in Chapter Five. Isobaths in that bay trend parallel to its shores, so that the ice-sea contact immediately prior to marine overlap would have been oriented from west to east off the southern shore of the Port au Port Peninsula, causing ice to swing from westwards to southwards in moving over it. On the eastern shore of the bay, the southwest-to-northwest trend of the ice-sea contact would have been similar to that of the glacier margin at the maximum, so that no such late-glacial change in flow direction occurred. The distribution of erratic till boulders along that coastline indicates only westward-flowing ice. Figure 26 shows the changing position of the ice margin and flow directions during the late Quaternary over southwestern Newfoundland. The regional picture of isochrones on ice margins in the Gulf of St. Lawrence is shown in Prest (1969), but those shown here have been derived from evidence gained in this study.

3. THE AGE OF THE EVENT AND ICE-CAP DIMENSIONS

Because it is the lowest and therefore the oldest unit of glacial

Figure 26


200 m
Laurentian Channel
100 m
200 m
100 m
14.5
14.0
13.7
13.5
13.
12.6
12.6
13.2
13.7
13.4
13.5
13.6
100 m
N
100 m
200 m
Speculative ice-marginal positions
during deglaciation of southwestern
Newfoundland and offshore zone
Ice-marginal position with date
in radiocarbon years X 10³
13.5 C¹⁴ date X 10³ years B.P.
Ice frontal "surge"
Land above 2,000 feet
16 8 0 16 mi

deposits, the age of the St. George's River Drift cannot be ascertained by bracketing it between dated deposits. The deposition of the unit preceded marine overlap in southwestern Newfoundland, so that the termination of the glacial phase may be dated at between 13,700 and 12,600 years B.P. in the area.

Two lines of evidence converge on the conclusion that this drift unit relates to the Wisconsin maximum, although no date for the beginning of the deposition can be obtained. Firstly, the glacial striae in the Port au Port area and erratic till boulder provenances indicate that the St. George's River Drift was deposited by ice flowing westwards off the highlands of southwestern Newfoundland. There is no evidence that more powerful ice flow from a Labrador centre affected the area. This suggests that the drift relates to the maximum extent of Wisconsin ice flowing from an island outflow centre.

Secondly, the till of the drift is a compacted basal till and it contains boulders of rocks which build the highest parts of the area. Ice, therefore, completely covered the land areas of southwestern Newfoundland during the deposition of the till and this would be most likely at the Wisconsin maximum. By 13,500 years B.P. the St. George's Bay shoreline was almost completely free of ice, and by 12,600 years B.P. the fjord arms of the Bay of Islands and Bonne Bay, and the open shorelines of the Gulf of St. Lawrence were deglaciated and ice was becoming more and more inactive. The deposition of the St. George's River Drift till, therefore, must have begun some time before these dates.

Little evidence is available which would enable the profile of the Newfoundland ice sheet at the glacial maximum to be estimated.

MacClintock and Twenhofel (1940) stated that, "Fresh erratics and roches moutonnees were found in abundance on enough critical upland summits to convince the authors that Wisconsin ice had completely glaciated the island" (p. 1775). I concur with this conclusion from an examination of summits in the Long Range and Bay of Islands Mountains. Atop Table Mountain, Bonne Bay, for example, fresh granite boulders can be seen easily amongst the frost-shattered debris of the dunite bedrock. Erratics also indicate that actively flowing ice crossed these summits.

At the glacial maximum the outer margin of the island ice cap would have been grounded on the gently-shelving sea floor off the west coast. It is not known how far out into the lowland of the Gulf of St. Lawrence the ice extended at this stage. The small-scale features of glacial erosion in the Port au Port area indicate strong flow westwards over that salient at the maximum, and the nature of the St. George's River Drift till suggests that ice had sufficient mass to render it extremely compact. I have assumed that the ice margin lay along the edge of the Laurentian Channel at 200 metres below present sea level, 40 kilometres (25 miles) west of the western extremity of the Port au Port Peninsula, for the sake of the computation made below.

Andrews (1968b) modified the formula of Weertman for the profile of a circular icecap. Weertman's formula states:

$$h^2 \simeq \frac{4\tau_{av}}{2\rho g}\,\ell - x$$

where

- h is elevation in metres on ice surface at x
- x is distance in kilometres from the centre
- τ_{av} is average shear stress in bars at the base of the ice
- ρ is density of the ice
- g is acceleration due to gravity
- ℓ is half-width of the ice cap

Andrews used average shear stress values of 0.5 and 1.0 bars to simplify the formula to:

$$h \approx 42D^{0.5} \text{ and } h \approx 85.^{0.5}\text{, respectively}$$

where h is height of ice surface at D, in metres
D is distance from margin of ice cap, in kilometres

Andrews cited studies by Hollin and Løken which gave values of the exponent in the equation of 0.57 and 0.58, and values of the varying "constant" from near 40 to near 100.

I have applied Andrews' simplification of Weertman's formula to western Newfoundland and have used a value of 0.57 for the exponent, and values of 40, 100, and the mean value of 70 for the "constant." Profiles derived from these computations are shown in figure 27. The highest point of land in Newfoundland, in the Lewis Hills, stands at 814 metres (2,673 feet) at a distance of 70 kilometres from the 200-metre isobath bordering the Laurentian Channel. The formula:

$$h = 100D^{0.57}\text{, where } D = 70 \text{ kms.,}$$

gives a value of 1,126 metres for the ice surface elevation above the 200-metre isobath. Since the 814-metre summit is 1,014 metres above that isobath, the summit lies beneath 112 metres (370 feet) of ice.

The "constant" could be adjusted to increase the thickness of ice above the summit, but it is apparent from these examples that the value of 100 is a minimum. The exponent is less adjustable since it shows a small variation in the studies quoted by Andrews (1968b) and is more closely related to the physical constants in Weertman's equation. The only possible source of the variation in the value of the constant is the average shear stress on the glacier bed. Since the ice caps to which the

Figure 27.

HYPOTHETICAL PROFILES OF WESTERN NEWFOUNDLAND ICE CAP AT WISCONSIN MAXIMUM

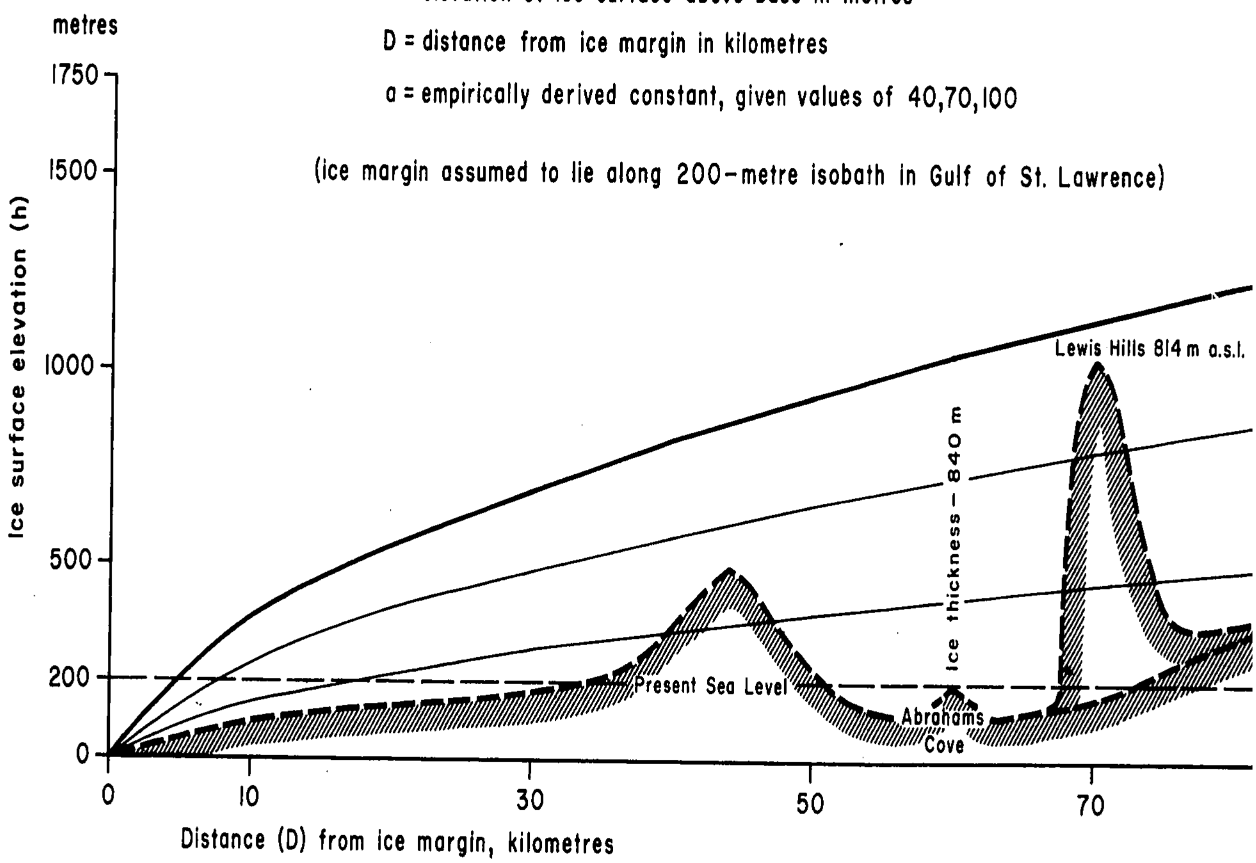
From equation $h = aD^{0.57}$,

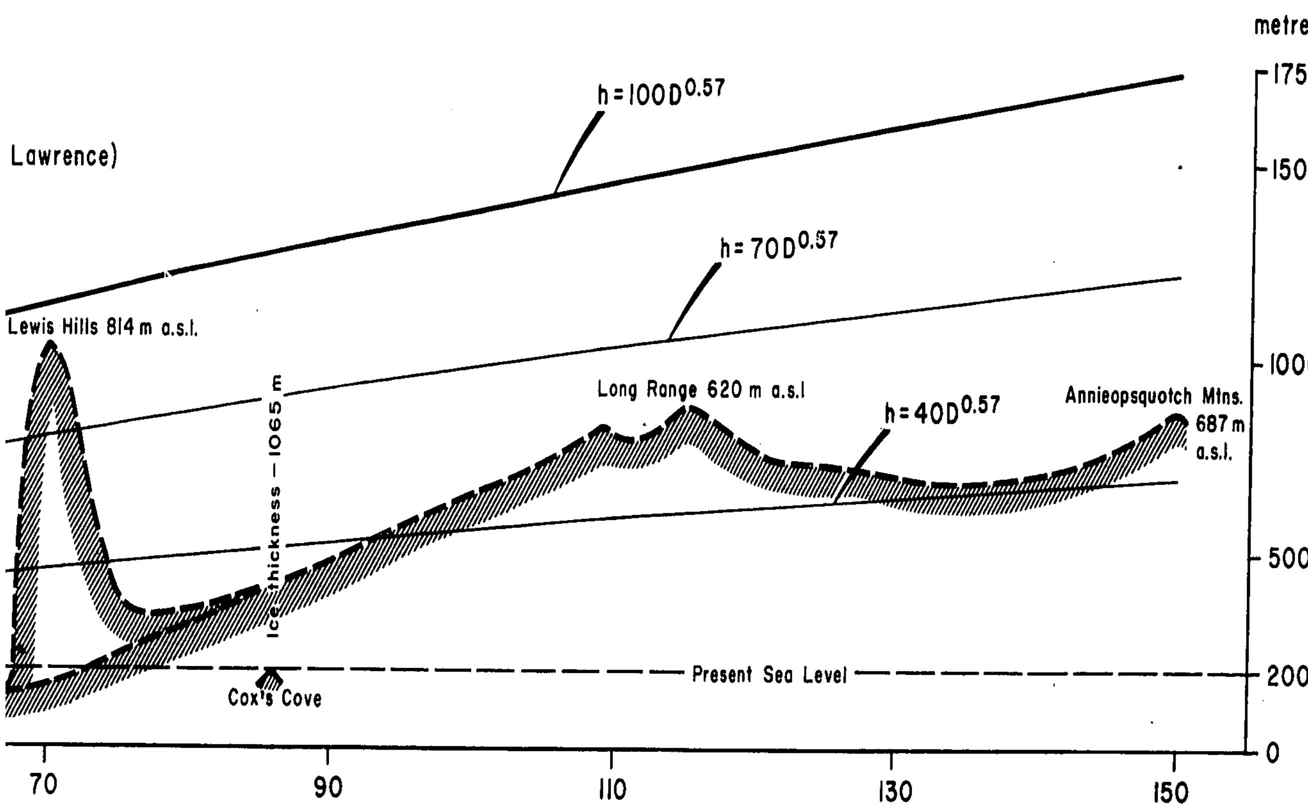
where h = elevation of ice surface above base in metres

D = distance from ice margin in kilometres

a = empirically derived constant, given values of 40,70,100

(ice margin assumed to lie along 200–metre isobath in Gulf of St. Lawrence)





metres
1750
1500
1000
500
200
0
h=100D$^{0.57}$
h=70D$^{0.57}$
h=40D$^{0.57}$
Lawrence)
Lewis Hills 814 m a.s.l.
Ice thickness — 1065 m
Long Range 620 m a.s.l
Annieopsquotch Mtns. 687 m a.s.l.
Present Sea Level
Cox's Cove
70
90
110
130
150

equation has formerly been applied are polar, the "constant" applicable to the Pleistocene temperate ice cap of western Newfoundland can be expected to be somewhat different. These computations also take no account of topographically induced changes in ice-surface slope over western Newfoundland where, with the high relief near the coastline, these would doubtless have occurred.

The values of ice thickness over points in western Newfoundland, computed above, are in agreement with those arrived at on the basis of topographic evidence. Fjord valleys have been eroded to at least 100 fathoms (approximately 200 metres) in the Bay of Islands and Bonne Bay. Immediately adjacent to these fjords, valley walls rise steeply to summits at near 800 metres. If the erratics on the summits are interpreted as evidence of flowing ice, ice thickness over the valley floors was over 1,000 metres.

The evidence of striae and other erosional micro-features of glacial origin indicates that, at the Wisconsin maximum, strong ice flow was towards the west and northwest from centres of outflow over the Long Range and Bay of Islands Mountains. The St. George's River Drift till is a lodgement till with no sign of any terminal features and was deposited by ice flowing towards the Gulf of St. Lawrence. The margin of this ice cap must be assumed to have lain in the Gulf of St. Lawrence. The ice surface profile tentatively applied to western Newfoundland indicates that, in order for the highest summits to have been covered by a minimum thickness of flowing ice, the margin of the ice cap must have stood close to the present 200-metre isobath in the Gulf of St. Lawrence, 35-60 kilometres (22-37 miles) offshore.

CHAPTER FIVE

Late-Glacial Marine Overlap and the Bay St. George Delta Event

1. THE EXTENT AND CHARACTER OF THE LANDFORMS AND DEPOSITS

MacClintock and Twenhofel (1940) gave the name "Bay St. George Delta" to a unit of glacio-marine deposits overlying the "St. George's River Drift" around the shores of St. George's Bay. Between Port au Port and Ship Cove, deltaic bottomset, foreset, and topset beds commonly comprise the bulk of deposits exposed in coastal cliffs (figure 28), and in several places along that coastline level cliff-top surfaces are fragments of this composite delta. The delta surface has in many places been terraced below the elevation of the upper marine limit or covered by a younger unit of glacial deposits. While fragments of this delta surface around St. George's Bay are the most prominent evidence for marine overlap, smaller delta terraces built in isolated coves and wave-washed till surfaces are widespread throughout southwestern Newfoundland along steeper and more rocky shores. Several different landform types are therefore associated with the late-glacial marine overlap, the "Bay St. George Delta" being merely the most prominent and continuous. I have referred to the Bay St. George Delta "event" to facilitate a comparison with the work of MacClintock and Twenhofel.

In this chapter only those deposits and landforms formed immediately following the onset of marine overlap will be discussed. The marine episode as a whole will be discussed in Chapter Seven. The onset of overlap was soon followed by another glacial event around St. George's

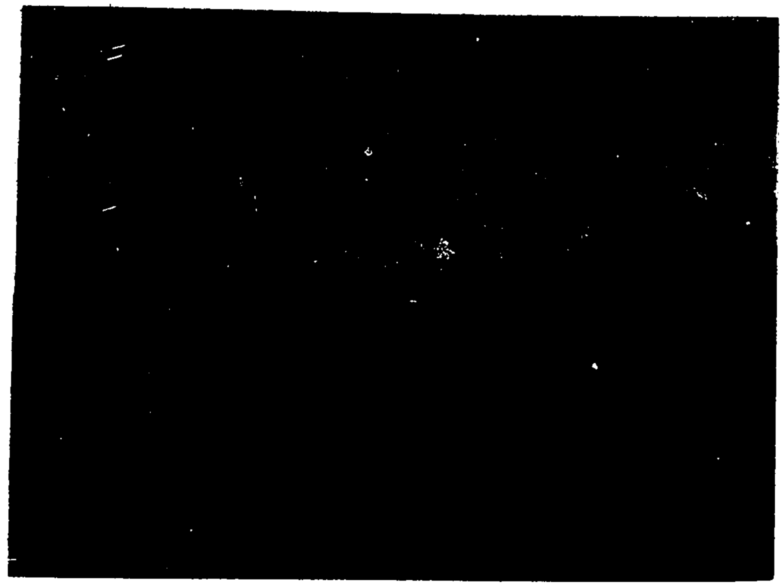


Figure 28: Coastal cliff section in foreset beds of Bay St. George Delta between Robinsons and Barachois rivers, St. George's Bay. Clay-silt bottom-set beds and till in cliff base.

Figure 29: Varved fine-grained sediments in cliff section near mouth of Fishels River, St. George's Bay.

Figure 28: Coastal cliff section in foreset beds of Bay St. George Delta between Robinsons and Barachois rivers, St. George's Bay. Clay-silt bottom-set beds and till in cliff base.

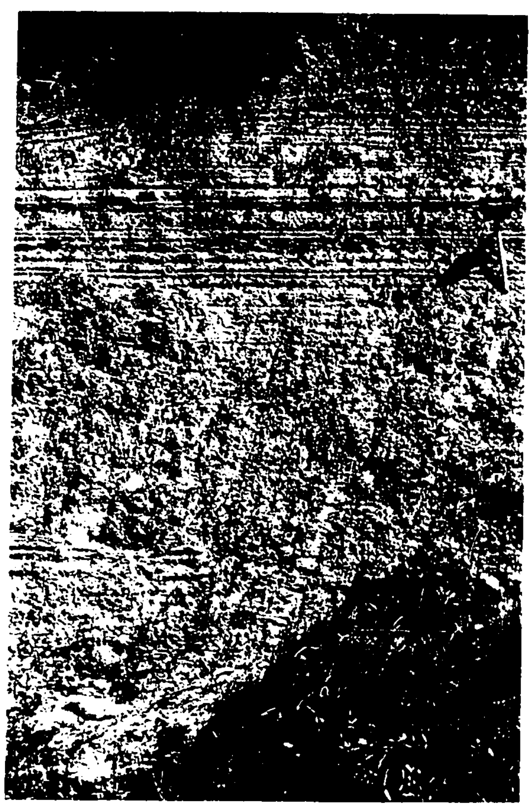


Figure 29: Varved fine-grained sediments in cliff section near mouth of Fishels River, St. George's Bay.

Bay and farther north, in the fjords, the progress of overlap was different again. Therefore, only the early part will be discussed here, and it will lead to a review of the later glacial event. Most of the interpretation of the evidence of marine events will be made in Chapter Seven.

The nature and extent of late- and postglacial marine landforms is shown in Plate VIII. Generally, the landforms can be classified into depositional and erosional types, the commonest examples of each type being the glacio-marine delta terrace and the wave-washed till surface, respectively. Upon the deglaciation of the present shoreline after the St. George's River Drift event, the only factors which caused one or the other of these types to form were the availability of glacial sediments immediately inland of the shoreline and of a stream to transport and deposit them into the sea. Where these conditions prevailed, deltas were built, but where they were absent marine waters merely washed the earlier till surface. In the latter case, the highest level of the sea can be detected in the junction of a smoother, washed till surface, commonly with large till boulders littering its surface, with a rougher, untouched till surface. This junction is often marked by a zone in which larger boulders are concentrated. This would be due to surf and sea-ice action which moved the boulders to the marine limit from the foreshore slope.

The factors responsible for the present-day distribution of marine limit landforms are, firstly, the preservation of the forms from destruction by later subaerial or marine erosion and, secondly, that they have not been obliterated by superincumbent younger deposits, primarily of glacial origin. In the area studied, the greater part of the shoreline was not affected by glacial deposition subsequent to marine overlap, so that marine landforms, where they were first formed, and have not since been eroded away, are still noticeable. Only along the St. George's Bay shoreline, between Port au Port and Ship Cove, near Highlands, has a subsequent glacial event caused the obliteration of most of the uppermost marine landforms. There, however, excellent stratigraphic evidence can be seen in the coastal cliffs that late-glacial marine waters were higher than levels at which marine landforms occur today. Before examining the elevation of the marine limit in southwestern Newfoundland, the nature of the glacio-marine deposits will be discussed.

Examination of many sections in the glacio-marine sequence demonstrates the similarity of depositional conditions over the entire study area. Above the St. George's River Drift surface (which may lie below the cliff foot) most commonly is seen a layer of massive, blue-black or red-brown, stony clay with a variable admixture of silt and sand. The thickness varies from one to fifteen feet. In some sections this basal clayey layer is overlain or replaced by a rhythmically stratified unit of silts and clays, suggestive of varve deposition. In the few localities where this occurs it is likely that large amounts of glacial meltwater were being brought into the late-glacial sea by a predecessor of one of the modern rivers draining to the Gulf of St. Lawrence. For example, near Fishels, a fifteen-foot thick sequence of varved silts and clays lies close to the present mouth of Fishels River (figure 29). Since the present river system draining into St. George's Bay was established as the ice-sheet wasted into the mountains, it can be reasoned that the glacial counterpart of Fishels River was contributing fresh meltwater and sediment to the late-glacial sea. Varve deposition cannot occur in water of average

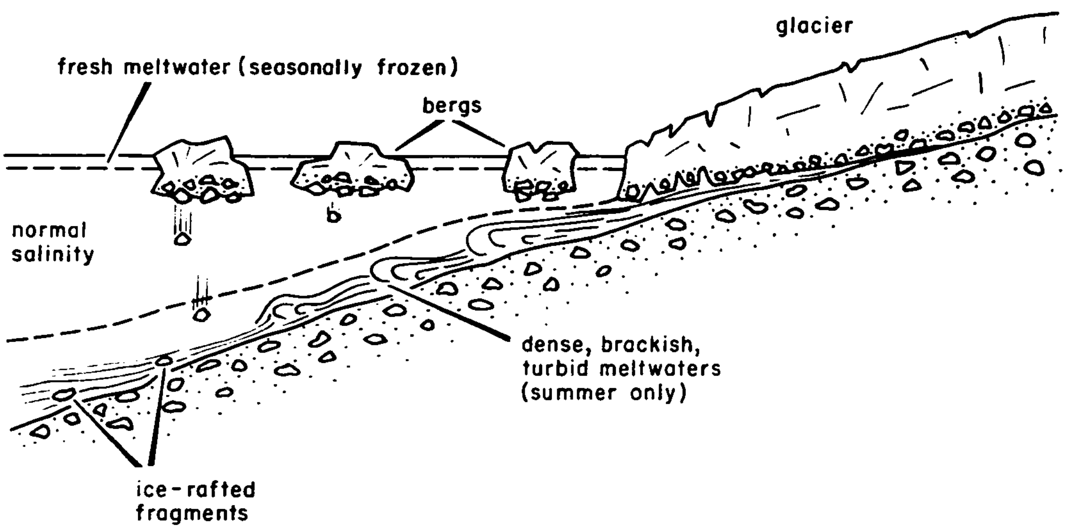


Figure 30. Offshore water density stratification and depositional environments in the early part of marine overlap in southwestern Newfoundland.

oceanic salinity since the negative charges on the faces of clay micelles become neutralized as cations in sea water are attached. The clay particles flocculate and sink to the sea floor together with the coarser particles. Sediment-charged meltwaters, however, would have reduced the salinity of inshore areas. Occurring in sudden pulses in a glacial environment, they would follow the bottom as density currents (figure 30). Kuenen (1951) states that such water would be even denser than sea water at the temperature of maximum density, 4°C. Brackish bottom waters would have deflocculated clay particles, allowing them more slowly to settle to the sea floor. A long season of sea-ice cover in late-glacial times would have reduced wave-generated turbulence at depth, contributing to the quiescent depositional environment necessary for varve formation.

In the massive clay layer, whether it be blue-black or red-brown in colour, many till pebbles and cobbles occur that are similar in lithology to the clasts of the St. George's River Drift till. They occur surrounded by unctuous clay and often the scars of barnacles are still visible on them (figure 31). Since barnacles are not tolerant of muddy bottom environments, it is possible that these boulders were dropped, along with the smaller gravel and sand in the clay, from the base of floating ice masses. In two sections examined in detail, at Tea Cove in Port au Port Bay and Benoit's Cove in Bay of Islands, the basal marine clays are so stony that, if it were not for the abundant and often frail molluscan shells in them, they could be mistaken for glacial till.

The junction of these fine-grained marine sediments with the underlying glacial deposits is marked by a transition zone, up to a foot thick, of coarse, clean sands, with included pebbles in some places

Figure 31: Sandy transition zone between St. George's River Drift till (below) and delta bottomset beds (above), Port au Port.

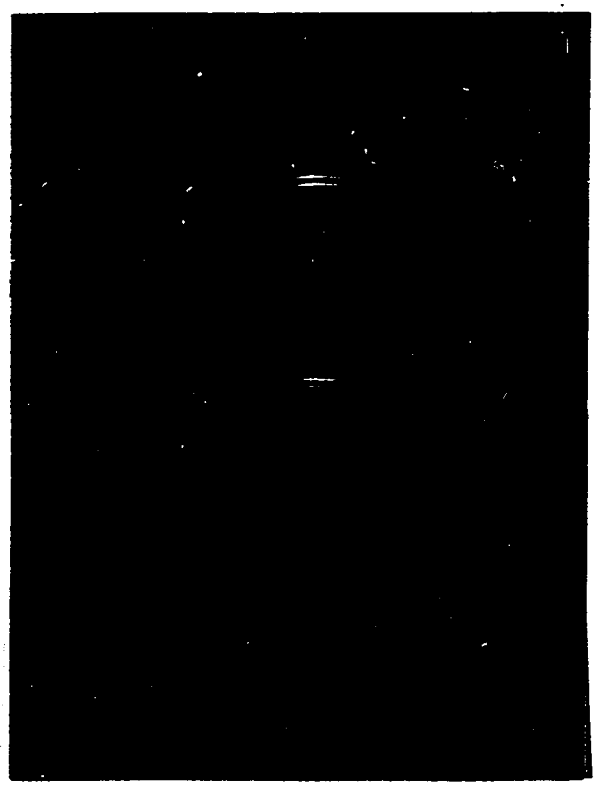


Figure 32: Junction of St. George's River Drift till and basal clays of marine sequence at Kippens, west of Stephenville. Note coarse transition layer.

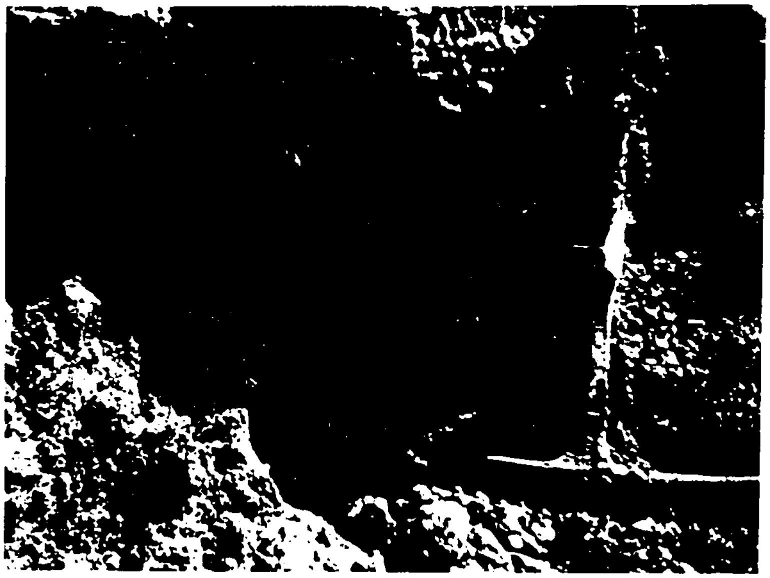


Figure 31: Sandy transition zone between St. George's River Drift till (below) and delta bottomset beds (above), Port au Port.

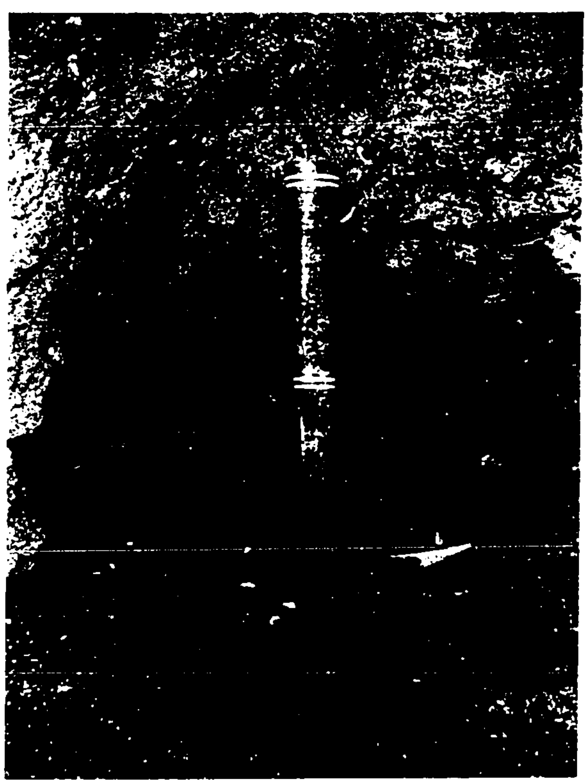


Figure 32: Junction of St. George's River Drift till and basal clays of marine sequence at Kippens, west of Stephenville. Note coarse transition layer.

(figure 32). This is suggestive of wave action on the glacial surface which removed the fines from the till. However, it will be shown in section two of this chapter that, upon deglaciation of present coastal areas, the sea was immediately as much as 200 feet deep over the surface of glacial deposits. The sandy transition zone between glacial and marine sediments must be the product either of wave-generated turbulence at depth or of erosion by density currents of sediment-charged meltwater.

It is in the massive clayey sediments, rather than the varved sequences, that the most abundant and varied marine shell fauna is found. Illustrations of the shell assemblages collected from several sites are found in Appendix A. Some points may be raised here relating to the environment of these shells as indicated by their state of preservation, thickness, and variety. The greatest variety and most complete state of preservation of the shells is found in the two sites mentioned above, at which the shells occur in a very stony, blue-black, unctuous clay. Barnacles are common at both sites, as are boulders with scars of these on one side of them. At Benoit's Cove, a complete test of Echinus sp. was exposed during the collection of other shells. Even the spines of the creature were found in the clay surrounding it. These phenomena can only be interpreted as indicating extremely quiescent bottom conditions in fairly deep water. At both sites, the clays occur at least 100 feet below the marine limit. An adequate explanation of the lack of turbidity on the sea floor can be given at each site, in terms of a cover of floating ice that remained for a short time after the onset of marine overlap because of local topographic conditions. These will be expanded upon when the radiocarbon dates are discussed later in this chapter.

Far fewer species, but larger numbers of individuals, have been recovered from other sites. At these, more turbid bottom conditions can be shown to have existed, usually by the coarser texture of the sediments. It is noteworthy that fewer species and indivuals are recovered from varved deposits. This may be due to the greater turbidity in such areas. Large influxes of glacial meltwaters from streams draining the ice-front would also have kept localized areas of the sea free of floating glacier ice and this is a possible explanation for the absence of till clasts in these varves, and of barnacles which would have colonized the under-surfaces of clasts as the latter emerged from the base of the ice. Further, in sandy deposits at the base of the marine sequence, species become fewer and shells thicker. Mya arenania, Mya truncata, and Macoma calcarea are the only species found in such environments and their shells are extremely sturdy.

The rhythmically bedded sediments which replace or overlie the massive unit have been referred to as varves, because of the regular alteration of thicker silty and sandy layers with thinner clayey layers. They can also be viewed as delta bottomset beds in an environment where stream and sediment discharge from land areas to the sea would have been markedly seasonal. The colour of these sediments always reflects the colour of the underlying till, across which streams would have flowed from the wasting glacier margin. The more massive clayey deposits rarely reflect this.

These fine-grained sub-units of the marine sequence pass abruptly up into a series of seaward-dipping delta foreset beds, which attains a thickness of 100 feet in places. In the St. George's Bay area, there are only two localities where the top surface of the delta remains intact, unaffected by post-formational terracing or obliteration by glacial deposition. These are at Port au Port (figure 33), and between Bank Head and Fishels River. Only there, therefore, can the coarse, horizontally-bedded gravels which overlie the foresets be called topset beds. Surface markings on the delta in the Bank Head area show the boggy floors of now-abandoned distributaries of the streams which built these deltas. In many other places around the St. George's Bay shoreline, delta foresets exposed in the cliffs are capped by similar coarse gravels, but it is known in these localities that the late-glacial sea once stood higher than the terrace surfaces. The latter, therefore, represent stillstands during the fall of sea level from the marine limit to its postglacial low position. One of these terraces is intimately associated with the deposits of the later, Robinsons Head Drift, glacial episode, so it will figure prominently in the discussion in Chapter Six.

MacClintock and Twenhofel (1940) commented upon the close relationship between the directions of dip of the foreset beds around St. George's Bay with the outlets of the present-day rivers. They concluded that the Bay St. George Delta was built by rivers which have not altered their courses very much in postglacial times.

Other, smaller, delta terraces in isolated coves in other parts of southwestern Newfoundland do not deserve detailed attention to the conditions of their formation. It remains to examine two other depositional features in connection with marine overlap at its highest stand. Harrys River enters the wide curve of the head of St. George's Bay from the northeast. Its source is close to the Bay of Islands, south of Corner Brook,

Figure 33: Raised delta at Port au Port, viewed towards the northeast, with Table Mountain in background.

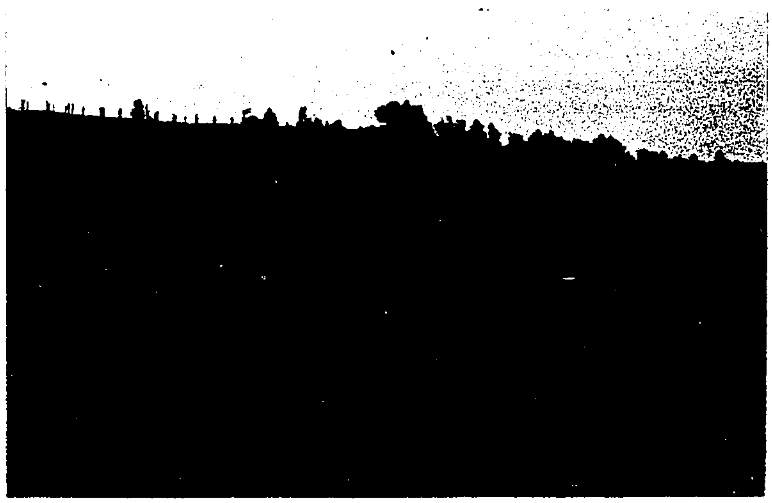


Figure 35: Section in gravels of raised beach ridges at Ship Cove, Port au Port Peninsula. Top of section about 30 feet above sea level.

Figure 33: Raised delta at Port au Port, viewed towards the northeast, with Table Mountain in background.

Figure 35: Section in gravels of raised beach ridges at Ship Cove, Port au Port Peninsula. Top of section about 30 feet above sea level.

and its large drainage area owes much to the evolution of the valley before glaciation. It flows in a broad lowland underlain by Ordovician clastic and carbonate rocks, flanked by the igneous body of the Lewis Hills to the west and the mass of the Long Range igneous and metamorphic complex to the east. In its middle and lower reaches it cuts through surficial deposits in wide meanders. Its lower reach is set in a modern flood plain from the west side of which a series of three wide terraces step up to the flanks of the Indian Head Range. These terraces, fragments of which are up to one mile wide and three miles long, were prograded into a sea which was intermittently lowered after the initial marine overlap. The highest rises to 120 feet at its proximal margin, yet it cannot be ascertained where upon its surface, or within its infrequently exposed deposits, the subaerial depositional environment gave way to the submarine. However, since the terrace could not have been built until ice had cleared the lower part of the valley, and late-glacial sea levels were high above present, this terrace must represent the marine limit. More will be said of the exact sequence of events in Chapter Six.

The last of the types of depositional landforms associated with marine overlap is the beach-ridged surface of a gently sloping, emerged sea floor. The best example of this is south of the mouth of Serpentine River, between the northwest edge of the Lewis Hills and the Gulf of St. Lawrence (figure 34). In this area a very gently sloping surface rises from the top of coastal cliffs 40-50 feet high, for two miles inland, in places to the base of a bluff at 130 feet, and in others to an impercep-tible break in the slope at a little above 160 feet. This surface is traversed by sub-parallel beach ridges, similar to those more recently

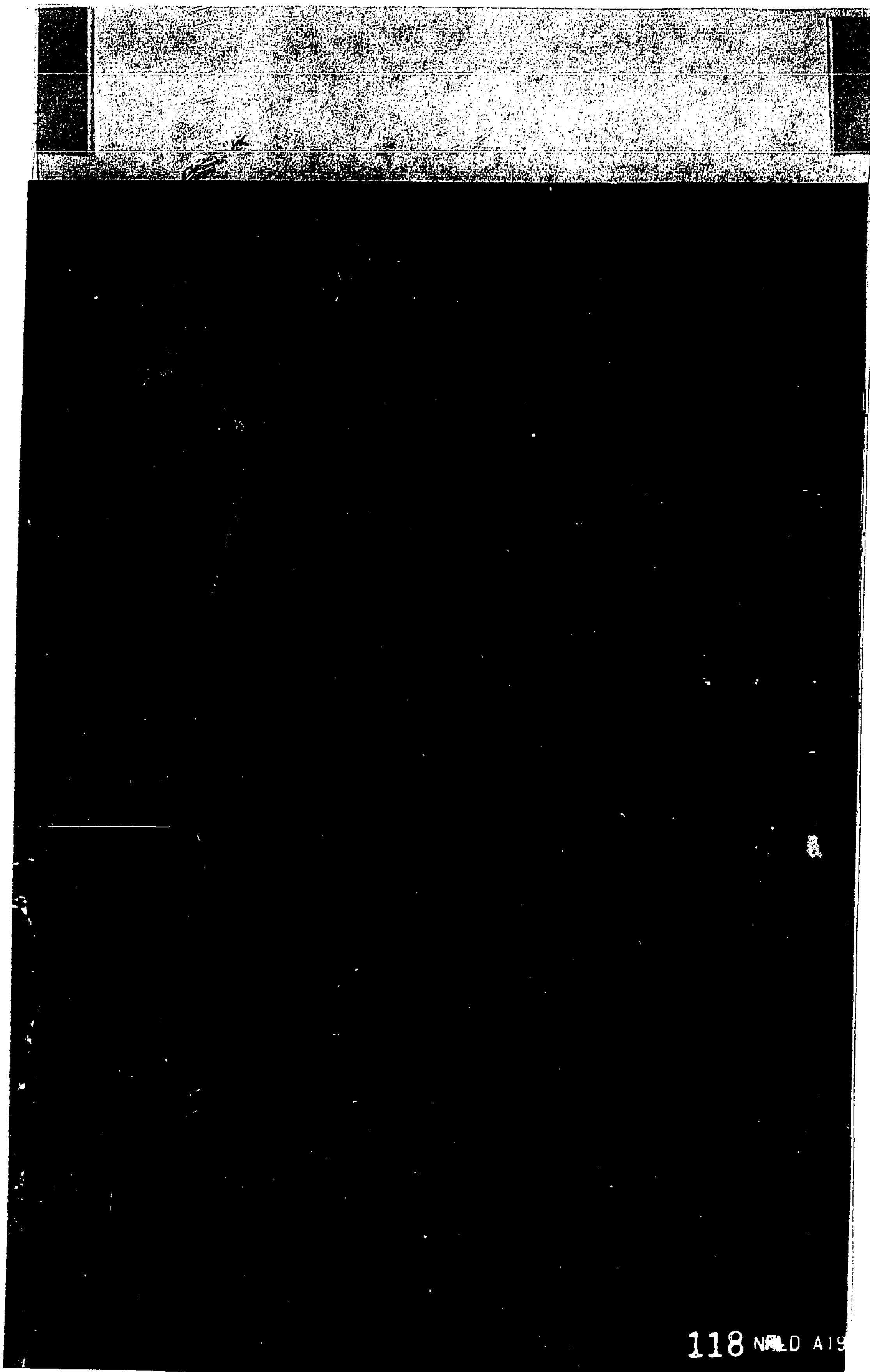
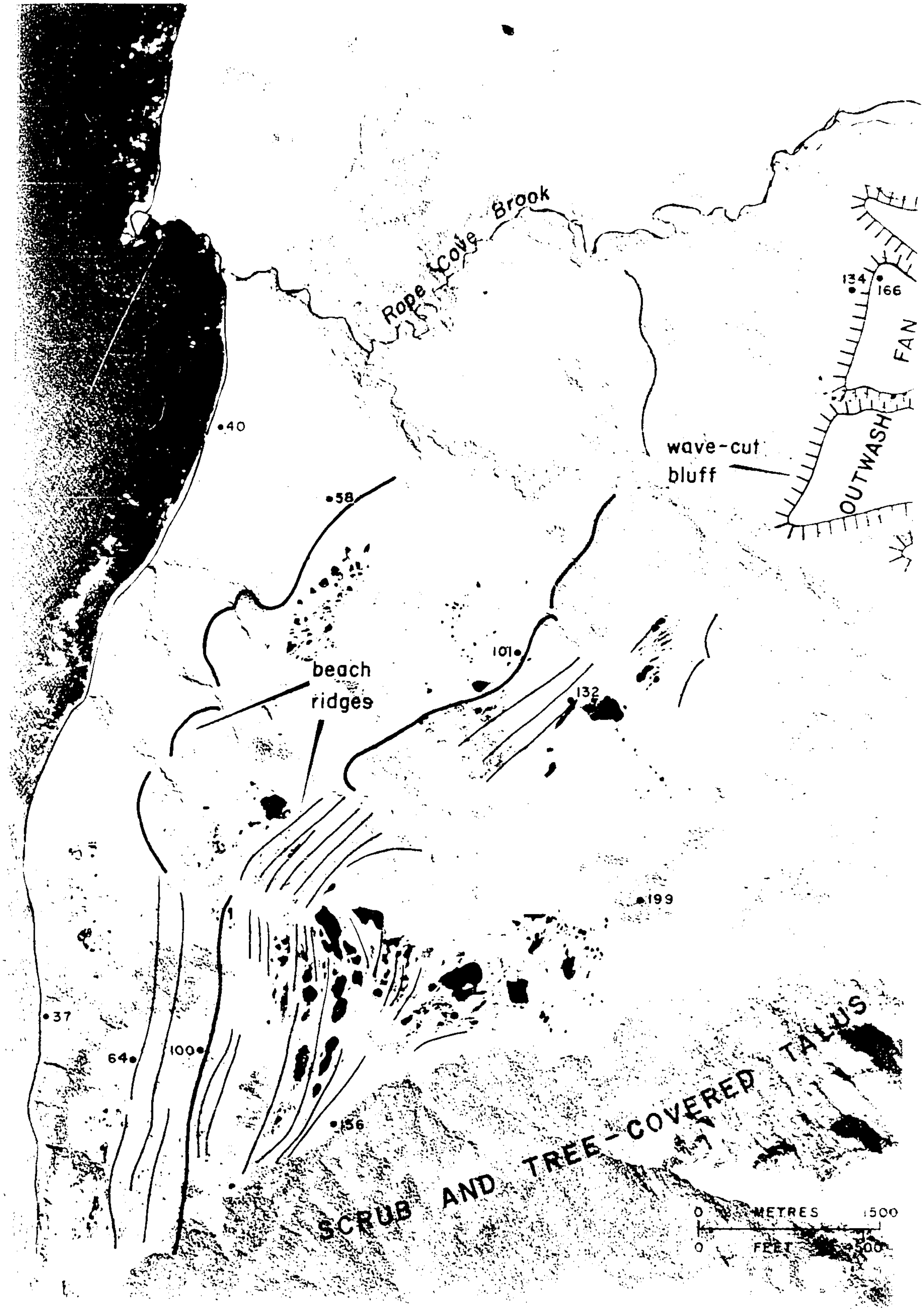


Figure 34: Raised marine features south of Serpentine River.


Rope Cove Brook
134
166
FAN
OUTWASH
wave-cut
bluff
40
58
101
132
beach
ridges
199
37
64
100
156
SCRUB AND TREE-COVERED TALUS
0 METRES 500
0 FEET 500

exposed around the borders of Hudson Bay. A gently sloping foreshore, plentiful supplies of bouldery debris, and a great distance of longshore fetch contributed to the building of these ridges. In a few other localities similar features are seen on a smaller scale. At Ship Cove, on the Port au Port Peninsula, for example, an apron of gravel, sloping regularly down to sea level from 50 feet, has well-developed, concave-seaward, beach ridges of two to three feet amplitude developed upon it (figure 35).

Erosional late-glacial marine landforms are, for the most part, restricted to wave-washed till surfaces. Important and problematical exceptions are found, such as the wave-cut rock platforms, and a fine example of a raised stack, at Trout River (figure 36). The wave-washed till surfaces are almost impossible to discern on aerial photographs and on the ground when the terrain is wooded. Since eroded bluffs at the inland margins of these surfaces are uncommon, their recognition is based upon the generally smoother and more level terrain below the marine limit; the more or less profuse litter of winnowed till boulders below the marine limit (figure 37); and the rare case of a concentration of till boulders at the marine limit, due to wave and sea-ice transport towards the former shoreline. None of these features is easy to identify unless the land has been cleared or is naturally covered with bog vegetation.

2. THE NATURE AND AGE OF MARINE OVERLAP

The elevation of all levelled marine limit features are shown on the map, Plate VIII[1]. Only the most general regularity of variation in

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1. In this study, elevations referring to specific raised marine features, or to former sea-level stands, are given in feet above higher

Figure 36: Wave-cut notch in metamorphic bedrock at back of "115-foot" delta terrace, Trout River.

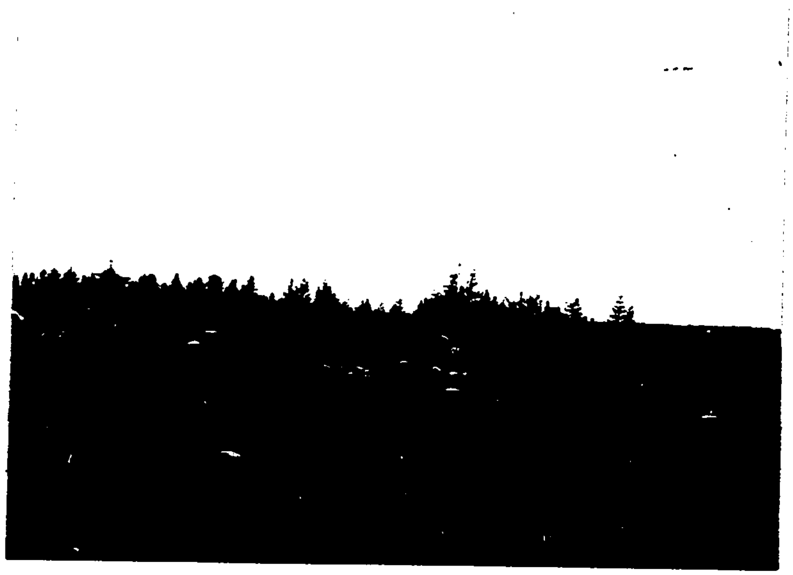

Figure 37: Wave-washed surface of St. George's River Drift till at St. George's. Highest point 80 feet above sea level.

Figure 36: Wave-cut notch in metamorphic bedrock at back of "115-foot" delta terrace, Trout River.

Figure 37: Wave-washed surface of St. George's River Drift till at St. George's. Highest point 80 feet above sea level.

their elevations is apparent. It can be seen that they are generally higher in the northern part of the area, but insufficient evidence is available to permit the reconstruction of isobases on the marine limit. This is most forcefully disallowed by the variation in the age of the marine limit over the area.

Fewer determinations of marine limit elevations were made in this study than in that of Flint (1940). However, features have been levelled in areas not covered by Flint, so that the spread of features is more even. Further, many features have been levelled in places where they cannot be fitted into the isobase pattern of Flint. Together with the radiocarbon age determinations of marine limit features, a reassessment of Flint's conclusions is given in Chapter Seven.

a. Radiocarbon-dated sites

Discussion of the nature of age of the marine limit will be organized around that of the dated sequences of marine deposits in southwestern Newfoundland. Seven radiocarbon dates became available between 1967 and 1969 on shells in deposits relating to the marine limit at six sites. Five of these are in the St. George's Bay-Port au Port Bay area and one is in the Bay of Islands. Data pertaining to the dates are shown in Table II.

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high water at large tides (HHWL). Features were levelled from Chart Datum (C.D.), using tide tables. Chart Datum is that level below which tides fall only one per cent of the time during the year, and is equivalent to lower low water at large tides (LLWL). Since evidence of marine activity will be found up to the higher high water mark, this datum has been used throughout. Elevations above Chart Datum have been lowered by six feet since this is the approximate average tidal range at large tides in the study area (Canadian Hydrographic Service, 1967).

TABLE II

RADIOCARBON DATES ON MARINE SHELLS FROM SOUTHWESTERN NEWFOUNDLAND

Locality	Lab. no.	Radiocarbon Age (years B.P.)	Elevation[1]		Marine Limit[1]	
			m	ft.	m	ft.
Abrahams Cove	G.S.C.- 968	13,600 ± 180	6	20	42.4	140
Abrahams Cove	G.S.C.-1074	13,700 ± 230	40.6	134	42.4	140
Port au Port	G.S.C.-1187	13,400 ± 290	1.8	6	34.2	113
Tea Cove	G.S.C.- 937	13,200 ± 220	1.8	6	30	100+[2]
Robinsons Head	G.S.C.-1200	13,500 ± 210	34.5	114	34.5	114+[2]
Highlands	G.S.C.- 598	13,420 ± 190	3.6	12	25.5	85+[2]
Cox's Cove	G.S.C.- 868	12,600 ± 170	34.5	114	48.5	160

[1]above HHWL
[2]The uncertainty of the elevation of the marine limit at Tea Cove, Robinsons Head, and Highlands is discussed in the text.

i. Abrahams Cove. The location of Abrahams Cove is shown in Plate VIII, and the dated stratigraphic sequence there is shown in cross-section in figure 38. The Quaternary deposits at Abrahams Cove partially fill a glacially eroded stream valley. The bedrock surface dips below sea level from 80 feet on each side of the cove. Where it is presently being stripped of a cover of St. George's River Drift till, the bedrock is seen to be smoothed, grooved and striated. The striation direction is shown on figure 23. The till is only four to six feet thick here and no ice-contact gravels are seen. The nature of the glacio-marine sequence is indicated in figure 38. Also shown is the relation of the position of the marine limit to the lower, regressional terrace surface. The date of 13,600 ± 180 years B.P. (G.S.C.-968) for the lower part of the marine sequence is on complete shells of Hiatella arctica. Specimens are shown in Appendix A. The shells are chalky, with patches of the outer skin, or periostracum and, in some cases, the ligament still intact. They occur in blue-black, clayey sands, with included till clasts. A date of 13,700 ± 230 years B.P. (G.S.C.-1074) was obtained on minute fragments of the shells of Hiatella arctica, found in the coarse, sandy matrix of an apron of angular limestone fragments. The shells occur within five feet of the marine limit, which lies at 140 feet above HHWL. The bouldery apron represents an accumulation of limestone fragments loosened by wave action from a bedrock cliff and incorporated in a matrix of coarse sand below the level of the sea. Similar accumulations are presently forming along the southern shore of the Port au Port Peninsula.

The comparability of the dates on shells which are separated vertically by 120 feet indicates that, immediately following deglaciation

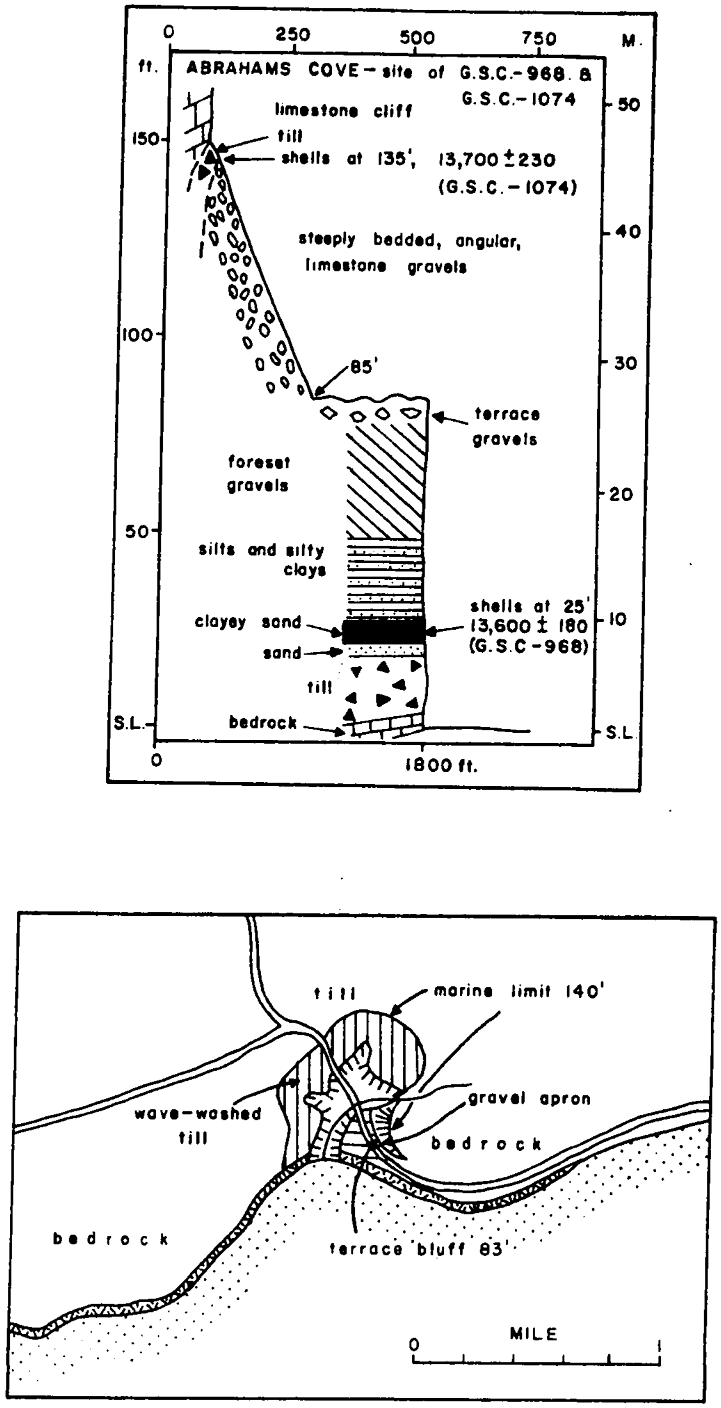


Figure 38. Quaternary stratigraphy and landforms, Abrahams Cove, St. Georges Bay.

of the cove, the sea was 120 feet deep over a steeply shelving shoreline. Bedrock cliffs rose above the marine limit at the sides of the cove, but at the head of this late-glacial embayment the limit was registered as a junction of a wave-washed till surface with a rougher till surface that stood above the sea. The similarity of the stratigraphic sequence at Abrahams Cove to that at other sites from which radiocarbon dates have been obtained on shells in the lower part of the marine deposits makes it possible to relate the marine limit at those sites to the dates obtained. Since the marine deposits belong to a penecontemporaneous wedge of sediment deposited during the deglaciation of the present shoreline, there is no possibility that deposits higher than those at the base are younger because of on-lap (i.e., transgression) or older because of off-lap (i.e., regression).

Depositional conditions following deglaciation are easier to understand if it is envisaged that, immediately prior to overlap, a eustatically rising sea level would be high against the front of an ice sheet which had depressed the crust beneath its weight. Isostatic recovery would be at its most rapid immediately before and after deglaciation of present coastal sites. Its rate would be at least three times as great as that of eustatic sea level rise. This has been shown, for example, in Arctic Canada (Andrews, 1968a), and at Boston (Kaye and Barghoorn, 1964). At the instant of deglaciation, the sea would register a marine limit feature against the newly-exposed land area, and its level would be falling rapidly as rebound continued.

Certainly, features that relate to stillstands in the regression of sea level will be seen cut into the first-formed marine features,

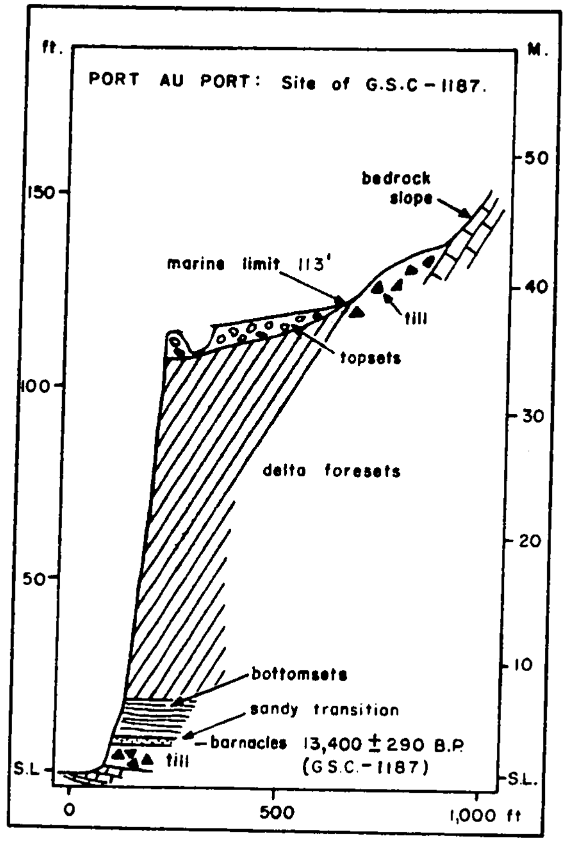
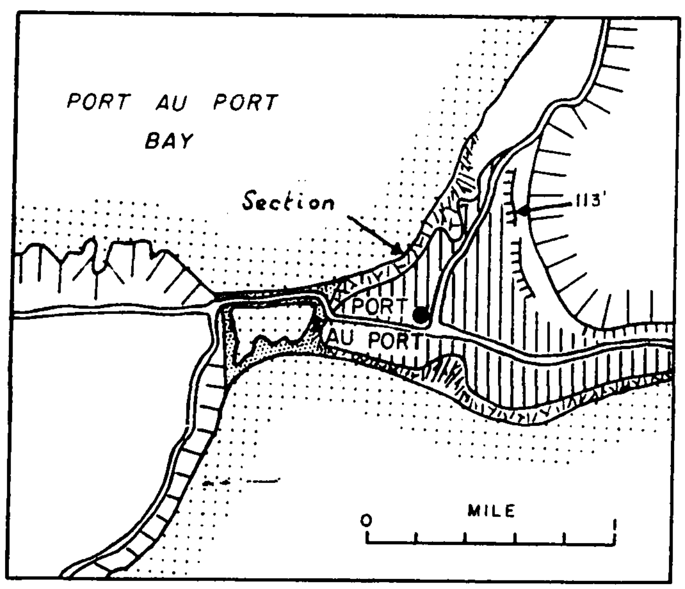


Figure 39. Quaternary stratigraphy and landforms, Port au Port.



whether the latter are erosional or constructional. At Abrahams Cove, for example, the back of a beach-ridged terrace surface stands at 80 feet above sea level as a regressional feature post-dating the marine limit.

ii. Port au Port. The location of the Port au Port site is shown on Plate VIII, and the sequence of deposits and landforms in figure 39. A radiocarbon date of 13,400 ± 290 years B.P. (G.S.C.-1187)[2] was obtained on barnacle shells collected from a zone of loose, bouldery sands between the top of the St. George's River Drift till and the bottom of red-brown, laminated bottomset beds of the Bay St. George Delta. The shells were found at five feet above mean high tide level, and once grew on the surface of boulders included in the sand. The scars of these barnacles are still plainly visible on the boulders (figure 31). The latter must, therefore, have been dropped from the base of floating glacier ice immediately after deglaciation of the narrows which once separated the Port au Port Peninsula from the main part of the island. Above the bottomset beds is a thick sequence of delta foreset sands and gravels, dipping southwest, that rises to a flat delta surface at about 110 feet above sea level. The surface of this raised delta is pitted by kettles above the Port au Port Bay and St. George's Bay shorelines. This indicates rapid glacio-marine deposition with ice blocks becoming incorporated into the sediment. The marine limit is a low, but prominent, bluff where the delta surface abuts against the rougher terrain of the St. George's River Drift (figure 40). It is at 113

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2. The shells were collected and submitted for dating by J. M. Shearer, Dalhousie University, Halifax, in the course of his own investigations on submarine deposits in Port au Port Bay. I would like to thank him for allowing me to include this information in this study.

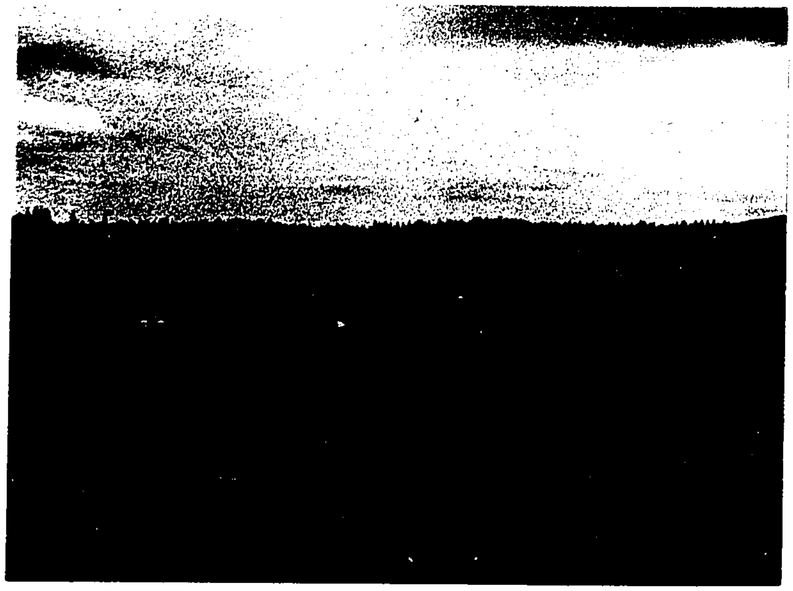


Figure 40: The marine limit at Port au Port; the junction between the rough till surface and the smooth raised delta surface is at 113 feet.

Figure 43: Robinsons Head from the air. Shell collection site in far left hand cliff face.

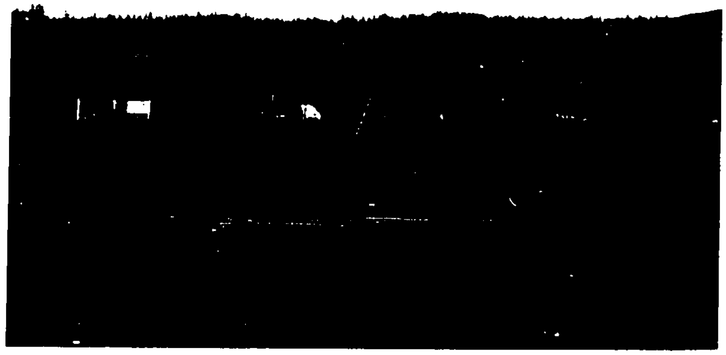


Figure 40: The marine limit at Port au Port; the junction between the rough till surface and the smooth raised delta surface is at 113 feet.

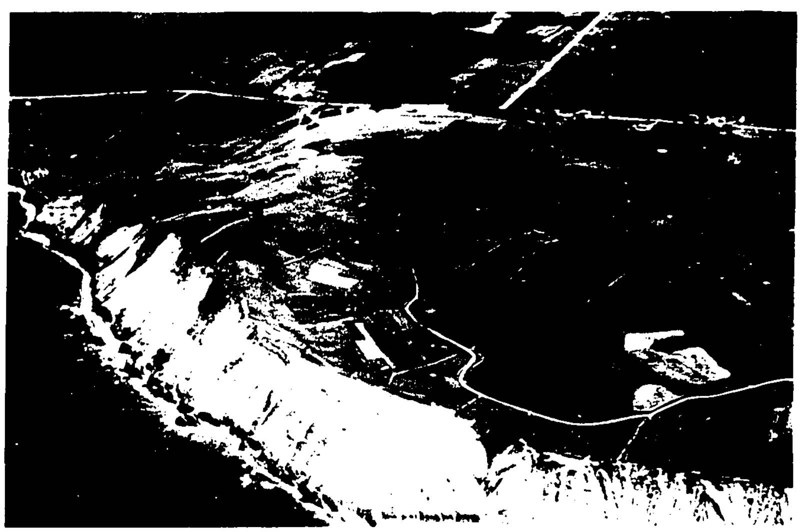


Figure 43: Robinsons Head from the air. Shell collection site in far left hand cliff face.

feet above HHWL. No regressional terraces are seen at Port au Port: if one or more were ever formed, they have been destroyed by cliff recession.

iii. Tea Cove. The location of Tea Cove is shown in Plate VIII, and the sequence of deposits and landforms in figure 41. As mentioned earlier in this chapter, Tea Cove is one of two sites investigated during this study which yielded a great variety of shell fauna. These are shown in Appendix A. A date of 13,200 ± 220 years B.P. (G.S.C.-937)[3] was obtained on a mixture of shells from a stony clay at six feet above HHWL. The clay overlies till of the St. George's River Drift and is overlain by horizontally-bedded marine gravels with an open fabric, which were deposited in relation to a sea level between 10 and 50 feet above present. These gravels are not related to the upper marine limit. The northern edge of this beach-ridged gravel surface is marked by its junction with a wave-washed till surface at 45 feet above HHWL. This boulder-strewn surface rises to about 70 feet near the coastal cliffs and to higher elevations inland, where it abuts against a bedrock cliff.

The marine limit at this site was not located in the field. The gently shelving surface west of Port au Port Bay was inundated to a greater inland extent than the more steeply sloping deglaciated surfaces at Abrahams Cove and Port au Port. The age of marine overlap here leads to the expectation of a marine limit close in elevation to that at Port au Port. But at this elevation the bedrock cliff bounding the low area inland

3. These shells were collected and submitted by J. M. Shearer, Dalhousie University. I gratefully acknowledge his permission to include the information in this study.

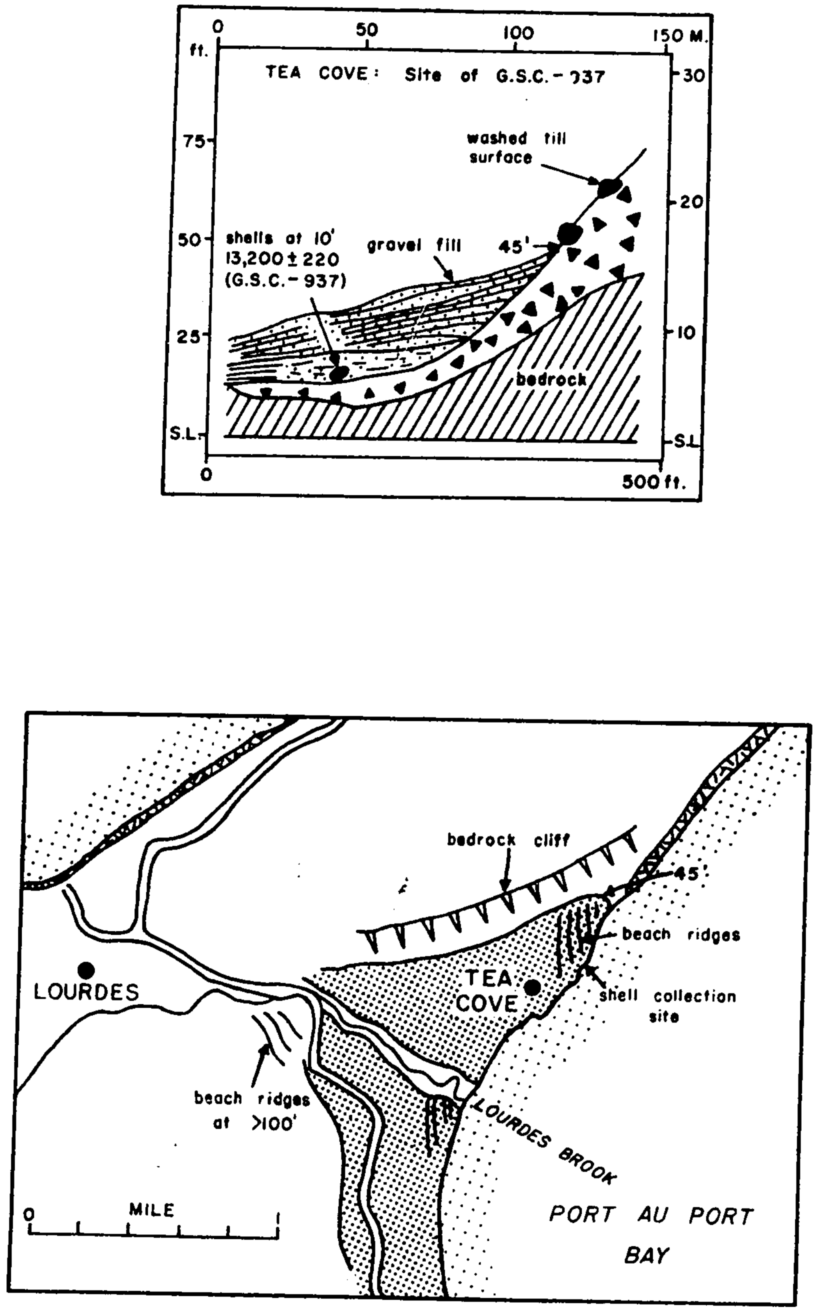


Figure 41. Quaternary stratigraphy and landforms, Tea Cove, Port au Port Bay.

of Tea Cove would have inhibited the erosion of a noticeable feature by a sea which was, moreover, being lowered by rapid isostatic rebound. Furthermore, the inclusion of an abundance of till materials in the basal part of the marine deposits indicates that glacier ice was afloat over the western shore of Port au Port Bay. This ice could even have blocked the ingress of the sea from the north across the northern end of the Long Point peninsula. Yet marine overlap at Tea Cove would still be possible via the low valley which forms a natural pass between Port au Port Bay and the open Gulf of St. Lawrence. The floor of this valley is below 50 feet. Whether overlap occurred through this valley alone or from both the west and the north, waves would be ineffective, for reasons of small fetch and floating ice masses, in producing noticeable landforms behind Tea Cove. On aerial photographs examined after the completion of field work, I noticed linear beach ridges on top of a till knoll on the south side of the valley mentioned above. The 1:50,000 topographic map shows these to be at just above and below 100 feet, so that it can be said that marine overlap occurred to at least that elevation. The terrace level at 45 feet is a regressional feature.

It is significant that Port au Port and Tea Cove are the only sites around the shores of Port au Port Bay where high marine limits, above 100 feet, have been found. These sites are the only ones at which marine overlap could have occurred from outside that bay: at Tea Cove from the valley to the west, and at Port au Port from the south, across the low area which is presently sealed off by a tombolo which joins the peninsula to the island. It is, therefore, reasonable to conclude that Port au Port Bay was ice-filled at 13,400-13,200 years ago, after the deglaciation of the St. George's Bay shoreline.

iv. Robinsons Head. The location of the Robinsons Head site is shown in Plate VIII, and the sequence of deposits and landforms in figure 42. A radiocarbon date of 13,500 ± 210 years B.P. (G.S.C.-1200) was obtained on fragments of shells of *Macoma calcarea* in a two-foot layer of brown, clayey silt at 114-116 feet above HHWL. The shelly layer overlies ice-contact gravels and till of the St. George's River Drift. These rest on Mississippian bedrock which rises higher in these cliffs than at any other site around St. George's Bay. Above the clayey silt layer is a unit of coarse, seaward-dipping sands, reminiscent of delta foreset beds, and these are overlain by twenty to thirty feet of coarse, cobbly kame gravel of the Robinsons Head Drift. The surface of this kame material rises along the cliff top to above 300 feet at the prominent summit of Robinsons Head (figure 43).

The shell-bearing clayey silt is not uniform in thickness or elevation along the cliff face in this headland. It rises and falls and thickens and thins as a reflection of the conditions under which it was formed. It is draped over the irregular surface of the glacial deposits beneath and thickens in the lower parts and thins over the rises. It is not certain to what elevation late-glacial sea water extended here. The 116-foot surface of the shelly layer at the collection site, which is the highest level it attains, must be regarded as a minimum since some depth of water is required for clayey silt to be deposited. The sands above, if they are deltaic, indicate marine activity to 140 feet above HHWL, and this fits the regional picture which will emerge as this discussion progresses.

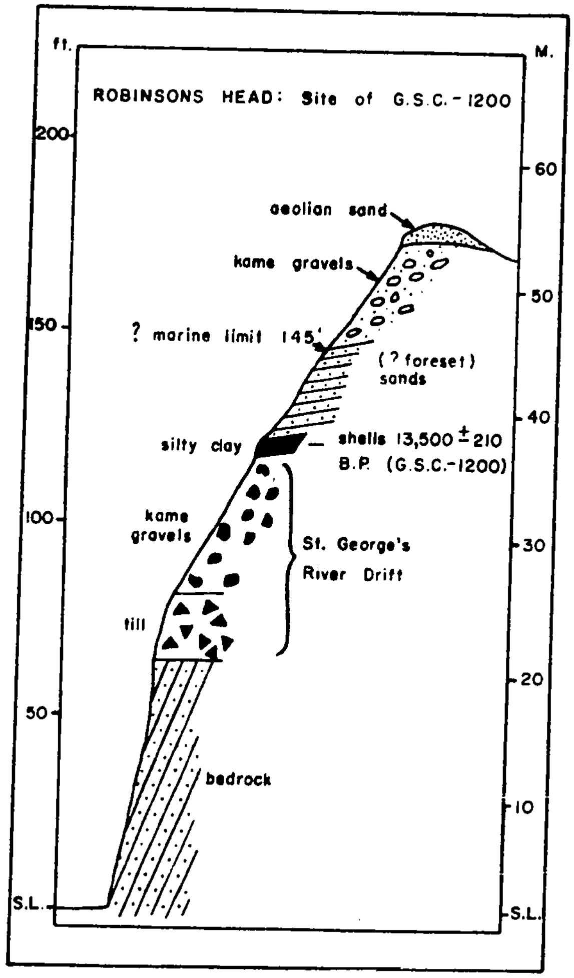
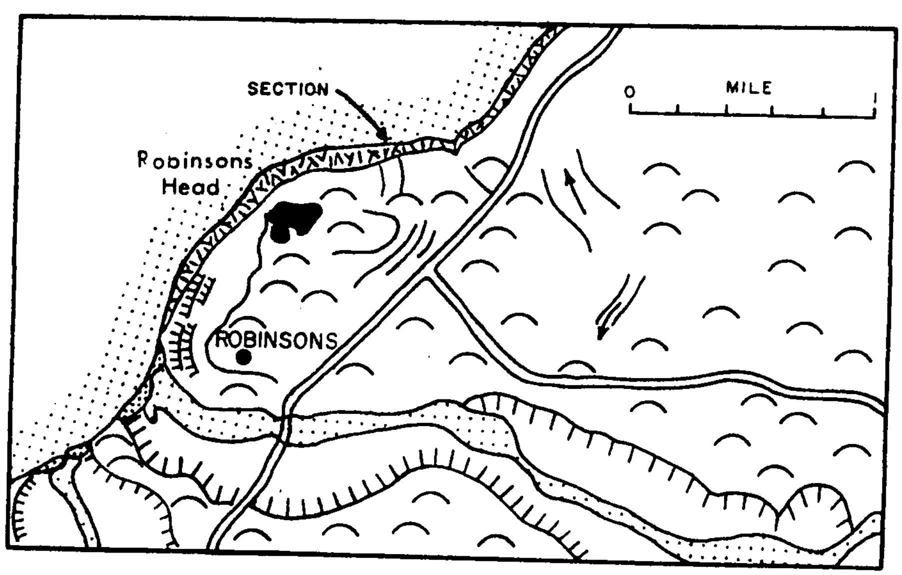
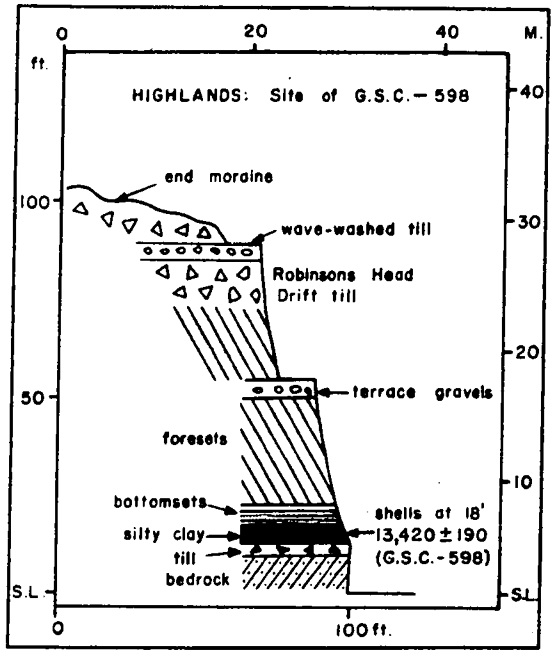


Figure 42. Quaternary stratigraphy and landforms, Robinsons Head, St. George's Bay.



The highest possibly marine landforms in the vicinity are at 120 feet, north of the headland, but strong bedrock support is offered there, so that the level nature of the surface may not represent marine bevelling. The highest definitely marine feature is south of Robinsons Head, near Jeffreys. There, the junction of the Robinsons Head Drift end moraine with a terraced delta surface is at 85 feet above sea level. This does not represent the marine limit, but the level to which the sea had fallen from the limit at the time the end moraine was being built. If the sands above the dated shells in the headland are taken as marine, a drop of sea level of at least 55 feet is indicated. At Robinsons Head, then, marine overlap and deposition of marine deposits was followed by at least a local readvance of ice into the sea when the latter had fallen below the marine limit.

v. Highlands. The location of the Highlands site is shown in Plate VIII, and the sequence of deposits and landforms in figure 44. A radiocarbon date of 13,420 ± 190 years B.P. (G.S.C.-598) (Lowdon and Blake, 1968) was obtained on fragments of shells of *Macoma sp.* in a blue-black clay, 12 feet above sea level. The clay overlies St. George's River Drift till and is overlain by silty bottomsets and sandy foreset beds of the Bay St. George Delta. At the collection site, the foreset sands are truncated by a regressional terrace surface at 50 feet above sea level. In a coastal cliff section at nearby Harbour Head, the foreset sands continue to an elevation of 70 feet, where they give way to a compact lodgement till which builds the upper part of the cliff to a height of 100 feet (figure 45). The cliff-top summit of Harbour Head is the seaward edge of a narrow lobe of end



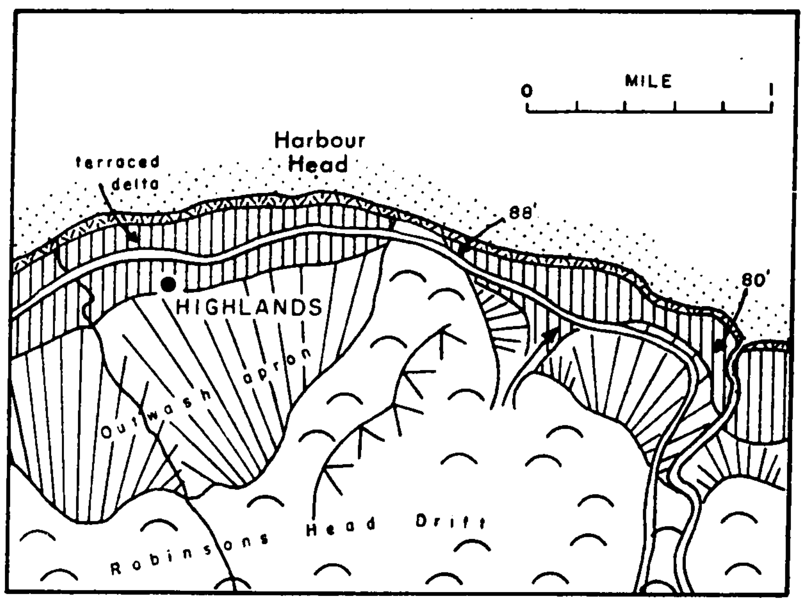


Figure 44. Quaternary stratigraphy and landforms, Highlands, St. George's Bay.

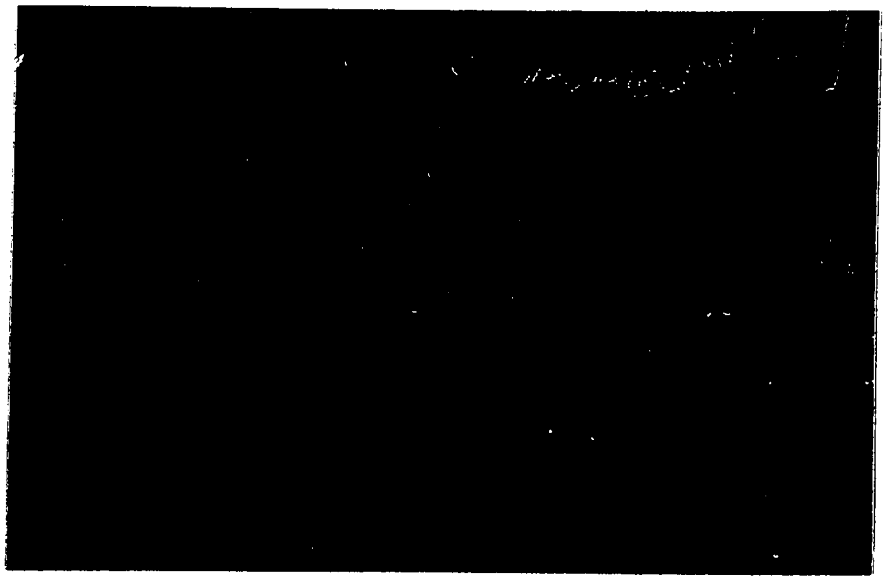


Figure 45: Lodgement till of Robinsons Head Drift overlying foreset beds of Bay St. George Delta at Harbour Head, near Highlands, St. George's Bay. Figure stands at junction.

Figure 45: Lodgement till of Robinsons Head Drift overlying foreset beds of Bay St. George Delta at Harbour Head, near Highlands, St. George's Bay. Figure stands at junction.

moraine of the Robinsons Head Drift. This lobe widens inland and merges with a sheet of ablation moraine. To each side of this summit an 88-foot terrace surface abuts against it and extends at approximately this elevation for several miles along the St. George's Bay shoreline towards St. Davids to the northeast and Ship Cove to the southwest. Beneath the surface of this terrace a zone of boulders can be seen above the lodgement till. They have evidently been winnowed from the till during the bevelling which produced the terrace. To each side of Harbour Head delta beds continue to the top of 90-foot cliffs.

A situation existed here which is comparable to that at Robinsons Head. Marine overlap occurred to an undetermined elevation. Deltaic deposition ensued, and a local readvance of ice into the sea occurred simultaneously with a stillstand at 88 feet, which bevelled the edge of the end moraine and terraced the surface of the pre-existing Bay St. George Delta.

vi. Cox's Cove. The location of Cox's Cove is shown in Plate VIII, and the sequence of deposits and landforms in figure 46. The sequence of deposits is exposed in two separate sites here; one along and just inland of the shoreline, and one about three quarters of a mile inland at a higher elevation. In the cliffs along the shoreline, sandy and silty clays with marine pelecypod and gastropod shells occupy the lower 25 feet, and are overlain by seaward-dipping sands and gravels which build a terrace surface at 70 feet. This terrace slopes up to a junction with a wave-washed till surface at 85 feet above HHWL about 100 yards inland. A regressional stillstand is indicated by this terrace.

About three quarters of a mile inland of the shoreline, in a

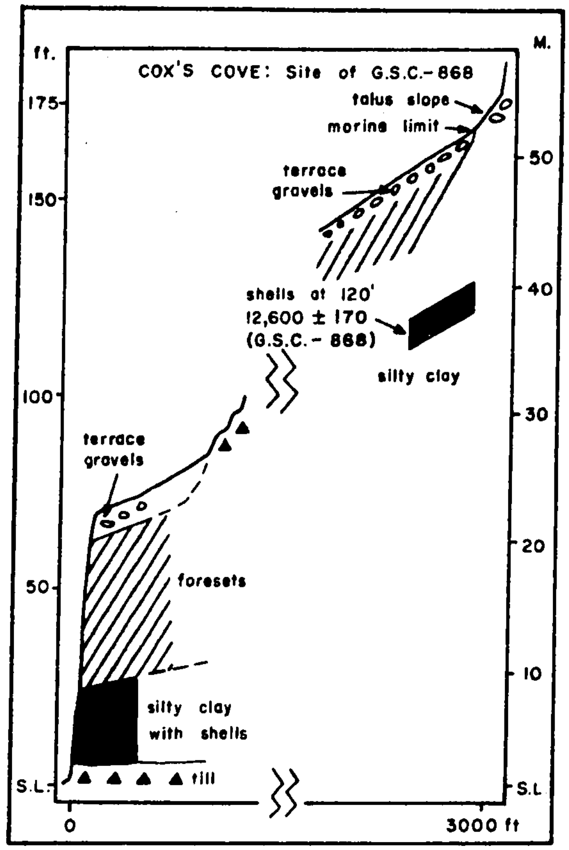
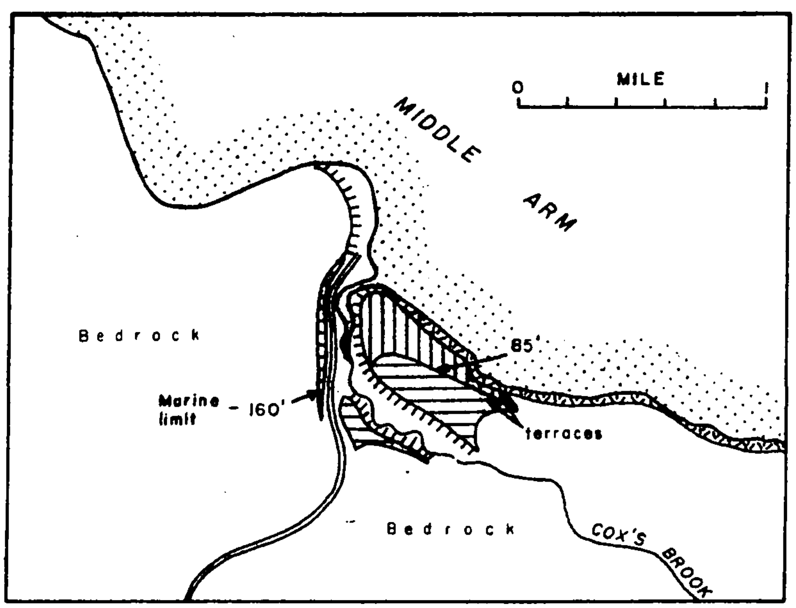


Figure 46. Quaternary stratigraphy and landforms, Cox's Cove, Bay of Islands.



roadside exposure, reddish-brown, laminated, silty clays outcrop to 120 feet above sea level. Shells collected from 114 feet have been radiocarbon-dated at 12,600 ± 170 years B.P. (G.S.C.-868). Above these clays seaward-dipping sands and gravels build a steeply sloping terrace surface which abuts against a talus slope at the foot of a bedrock wall at 160 feet. This junction marks the upper marine limit.

b. Other sites

Two more sites are of particular interest in the treatment of marine overlap given in this chapter. Radiocarbon dates are not available for the marine phase at these places, but a reasoned consideration of their topographic situation and the elevation of the marine limits, leads to conclusions which extend the reconstruction of the chronology of marine overlap to a wider area.

At Humbermouth, at the head of Humber Arm, Bay of Islands, a delta terrace fragment remains on the south side of the arm. A vertical face, presently being worked for construction materials, overlooking the arm was cut by the marine erosion of a delta that was built across the mouth of the Humber River immediately following deglaciation of the Humber Arm fjord. Another, more extensive, remnant of this delta is found across the arm from Corner Brook, upon which the city cemetery has been established. The lower Humber delta terrace was built in contact with active glacier ice at its proximal margin. In the working face of the pit in this terrace, seaward-dipping foreset beds can be traced up-dip to a junction with a massive, compact lodgement till. The till is a product of a glacier tongue occupying the lower Humber valley at the time of marine overlap at the head of Humber Arm. The ice-contact face is shown on Plate VIII.

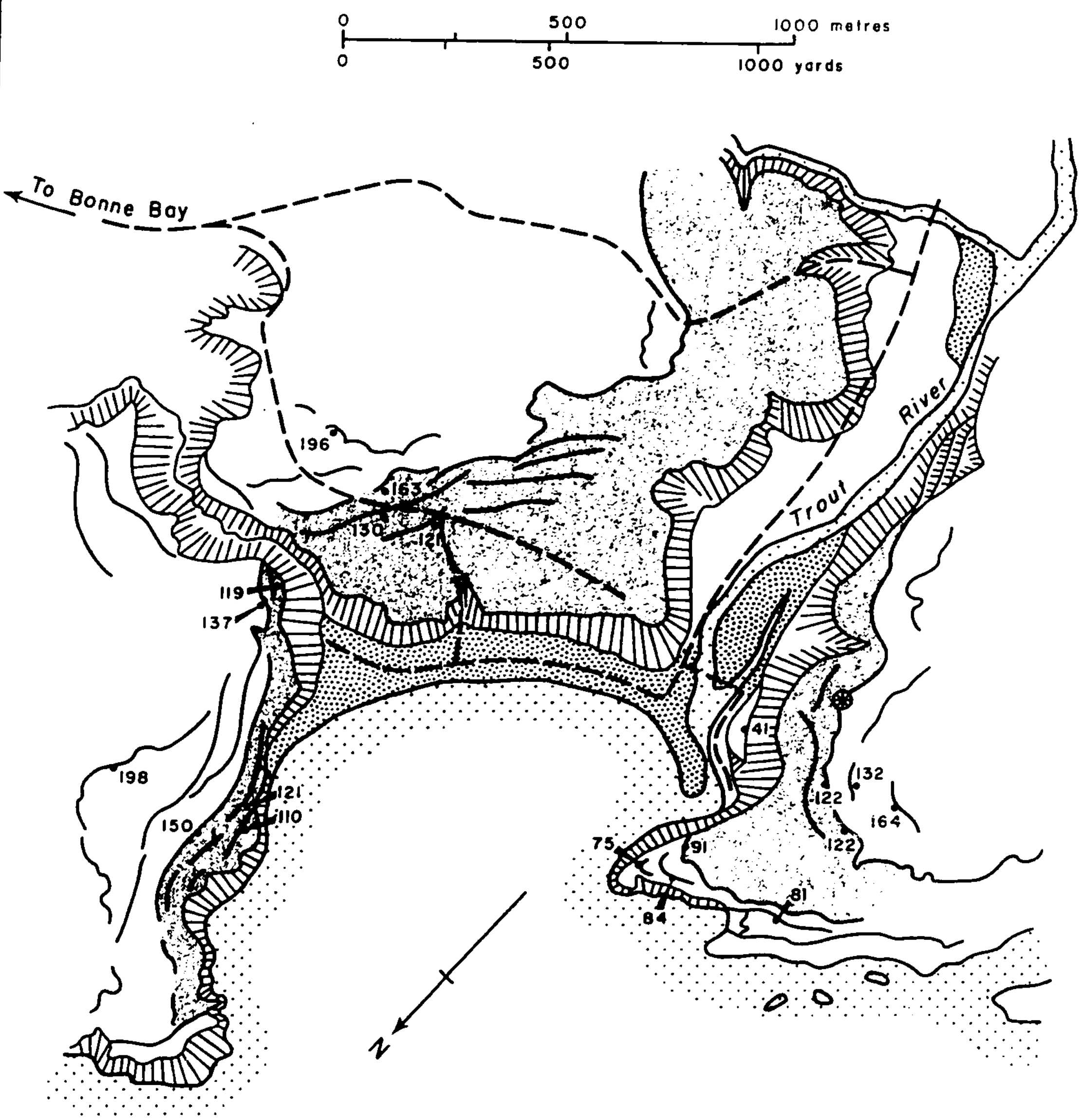
The top of the foreset beds in this delta lies at 158 feet above HHWL and the topsets are at 163-164 feet. The fjord-side locations and the comparable elevations of the terraces at Cox's Cove and Humbermouth may be taken as indicating their comparable age, i.e., 12,600 years B.P.

At Trout River, on the open Gulf of St. Lawrence coast, between the mouths of the Bay of Islands and Bonne Bay (Plate VIII), is seen the most spectacular series of marine terraces to be seen in at least south-western, if not the entire area of, Newfoundland. Figure 47 shows the relation of the terrace landforms to the surrounding rugged topography. The upper marine limit is marked by a junction of washed and untouched till surfaces at about 195 feet around the semi-circular cove. Faint traces of similar features were seen to about 230 feet above sea level, but their positive identification as being of marine origin was not possible, so they have been omitted here. Several low bluffs mark minor regressional stillstands below the marine limit, between it and the major constructional terrace feature at 114-117 feet. Below this latter feature marine levels are seen in terraces cut into the sediments of this delta, and in minor wave-cut platforms cut across bedrock on the headlands that project across the mouth of the cove. While no dates are available on marine overlap at Trout River, due to an absence of shell fauna, the surrounding topography provides evidence that the onset of marine conditions was delayed past the time at which it occurred around more gently shelving shorelines, such as St. George's Bay.

The cove at Trout River lies at the seaward end of an immense glaciated trough cut through the Bay of Islands Mountains, which provide

Figure 47.

TROUT RIVER



LEGEND

- Composite delta terrace
- Raised shoreline traces
- •130 Elevation of marine feature above mean low tide
- Coastal and riverine gravels
- Prominent bluff
- Raised sea stack on "115-foot" terrace
- Road

a backdrop to the cove. Near-vertical walls plunge 2,000 feet from plateau surfaces to the level of two ponds in the trough; one nine miles long, separated by a low bar from another six miles in length (figure seven). These ponds have not been sounded, so the depth of glacial erosion is not known, but the surface of the ponds is at sea level, so the trough may well be a fjord. This topographic situation would be conducive to the waning of ice from the shoreline at the same time as in the situations encountered in the Bay of Islands.

Marine overlap was not immediately marked by delta building at Trout River. Some 80 feet of relative sea level fall occurred before conditions conducive to delta building prevailed. This may be explained in one of two ways. Firstly, a glacier tongue in the Trout River canyon may have been debouching its meltwater southwestwards down the valley that leads to the sea at Chimney Cove before it poured into the sea at Trout River. Secondly, there may not have been glacial meltwater available from the active ice tongue in the canyon until some time after marine overlap. Insufficient evidence is available to permit a more positive explanation.

Two features of the marine landforms at Trout River mark them off from those seen everywhere else in the study area. One is the presence of a 50-foot high "fossil" sea stack and of wave-cut notches at the back of the 115-foot delta terrace on the southern side of the cove. Another is the flight of wave-cut platforms cut in resistant bedrock on the headlands at the mouth of the cove. Both sets of features seem to indicate either more prolonged stillstands of sea level than are evident in other areas, or more powerful wave energy on this open coastline. The question of regressional stillstands will be discussed in Chapter Seven, within the general topic of changes of land and sea level. It is certainly possible that more powerful wave energy can be built up along this, than along a more gently shelving, shoreline. Considerable fetch is available in a northwesterly and westerly direction and frictional drag on the bottom would not be effective until wave trains were close to the land. Little more of value can be added here, however, since the investigation did not proceed in this direction.

With this review of the nature and age of late-glacial marine overlap, a later glacial event must be dealt with in the next chapter, before proceeding to an integrative interpretation of the entire marine episode.

CHAPTER SIX

The Robinsons Head Drift Event

1. THE EXTENT AND CHARACTER OF THE LANDFORMS AND DEPOSITS: A GENERAL VIEW

MacClintock and Twenhofel (1940) gave the name "Robinsons Head Drift" to a sheet of glacial deposits spread widely over the hinterland of St. George's Bay. The name recognized the scenic prominence of that headland on the eastern shore of the bay. These authors mapped the extent of the deposits with remarkable accuracy (Plate IV). The only doubt arose in the Harrys River lowland, but their tentative projection of the boundary between better known localities has been confirmed in this study. They showed that the lobate margin of the drift sheet was intersected by coastal cliffs along the St. George's Bay shoreline, and that the drift does not occur west of the village of Port au Port.

MacClintock and Twenhofel, and Flint (1940), concluded that the ice sheet which produced the Robinsons Head Drift moved across a shallow sea floor;

> The intimate relationship, in some of the coastal exposures, between till of this drift and marine deposits suggests that the ice advanced into shallow sea water. (MacClintock and Twenhofel, 1940, p. 1753.)

> At several points along the southeast coast of St. George's Bay, marine sediments occur interbedded with locally thick layers of till, the base of the highest till zone in this relationship lying at an elevation of 40 feet . . . the relations clearly indicate active glacier ice moving over a shallow sea floor. (Flint, 1940, p. 1776.)

The most widespread landform within the drift sheet is developed on ablation till which mantles the foreland west of the Long Range Mountains and south of the Lewis Hills. The surface of the till is low and hummocky, with a few wide, flat, boggy areas. In only a few localities was evidence of flow moulding of the till by ice seen (see Plate VIII). Generally, the till appears to have been let down on to the surface as the ice sheet stagnated. Small areas of rough kame and kettle terrain within the drift border attest to local stagnation. In areas closer to the major streams draining the foreland, deposits are noticeably more stratified than where they can be seen on interfluves. This suggests the presence of abundant meltwater concentrated along the lines followed by present streams. In topographically favoured situations, eskers occur in this drift. The complex esker system in the Harrys River lowland shows magnificent examples. There, they appear to be related to the large amounts of meltwater pouring from the western end of Grand Lake, once that gorge had been produced by watershed breaching, and from the flanks of the Indian Head Range. Stagnant ice conditions are necessary for esker formation, so these are late features. These eskers occur at the break in slope between the Indian Head Range and the Harrys River lowland, where meltwaters would have been concentrated. They also lead into two gorges where sediment-charged meltwaters played a role in watershed breaching across the north-south axis of the Indian Head Range (see Chapter Three).

The most prominent landform in this drift sheet, however, is the kame and kettle moraine which marks its outer limit. In localities such as Robinsons Head and Bank Head, kame summits stand up to 300 feet above adjacent morainic and terrace surfaces (figures 43 and 48). The moraine is most prominent at the distal margins of lobate projections of the periphery of the drift. Between these lobes the moraine surface grades

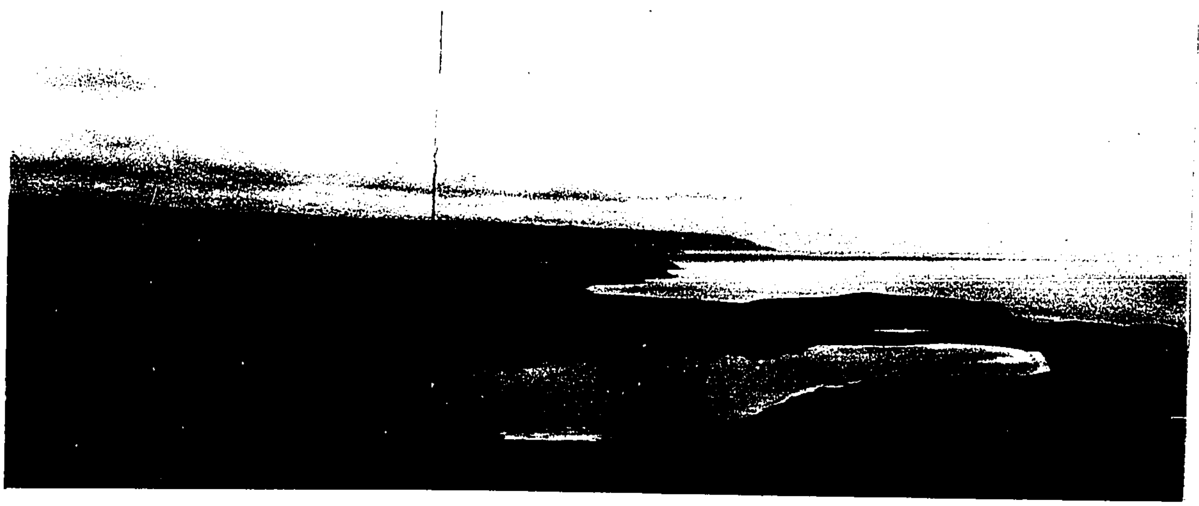


Figure 48: Kame-and-kettle terrain of Robinsons Head Drift end moraine at Robinsons Head, viewed towards the south from summit at near 350 feet. Cape Anguille Mountain on skyline.

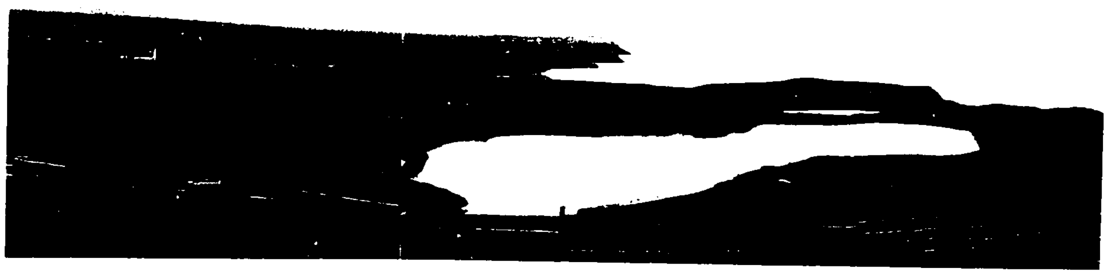


Figure 48: Kame-and-kettle terrain of Robinsons Head Drift end moraine at Robinsons Head, viewed towards the south from summit at near 350 feet. Cape Anguille Mountain on skyline.

indistinctly into the surfaces of raised deltaic terraces.

The nature of the Robinsons Head Drift sediments varies according to processes of their formation and to the source of the deposits. The ablation till is notably coarse, as shown in curves III, IV, and V in figure 49. The Irishman's Brook sample (curve II) was taken from the coastal cliffs near Bank Head below the marine limit, and its high silt content betrays its subaqueous origin. The Harbour Head sample (curve I) is a lodgement till taken from the top of the cliff at Highlands, described in section two, Chapter Five. Both these localities show evidence that ice which deposited these tills moved into the sea across sandy marine sediments.

Elsewhere in southwestern Newfoundland, outside the St. George's Bay hinterland, the surface is underlain by a mantle of till which relates to the wastage of ice masses into the mountain areas. In places, hummocky end moraines are visible adjacent to areas of outwash deposited close to the level of a regressing late-glacial sea. The relation of these moraines to that of the Robinsons Head Drift in the St. George's Bay area will be discussed after the treatment of features seen around that bay.

2. THE RELATION OF THE EVENT TO MARINE OVERLAP

Around St. George's Bay, the margin of the Robinsons Head Drift lies close to the shoreline and in several places the cliffs are cut back into salient lobes of hummocky end moraine. It is convenient to treat the evidence of the sequence of events associated with the formation of the moraine with reference to three areas: the Harbour Head to Bank Head section, the Harrys River lowland, and the Harmon Field to Port au Port section.

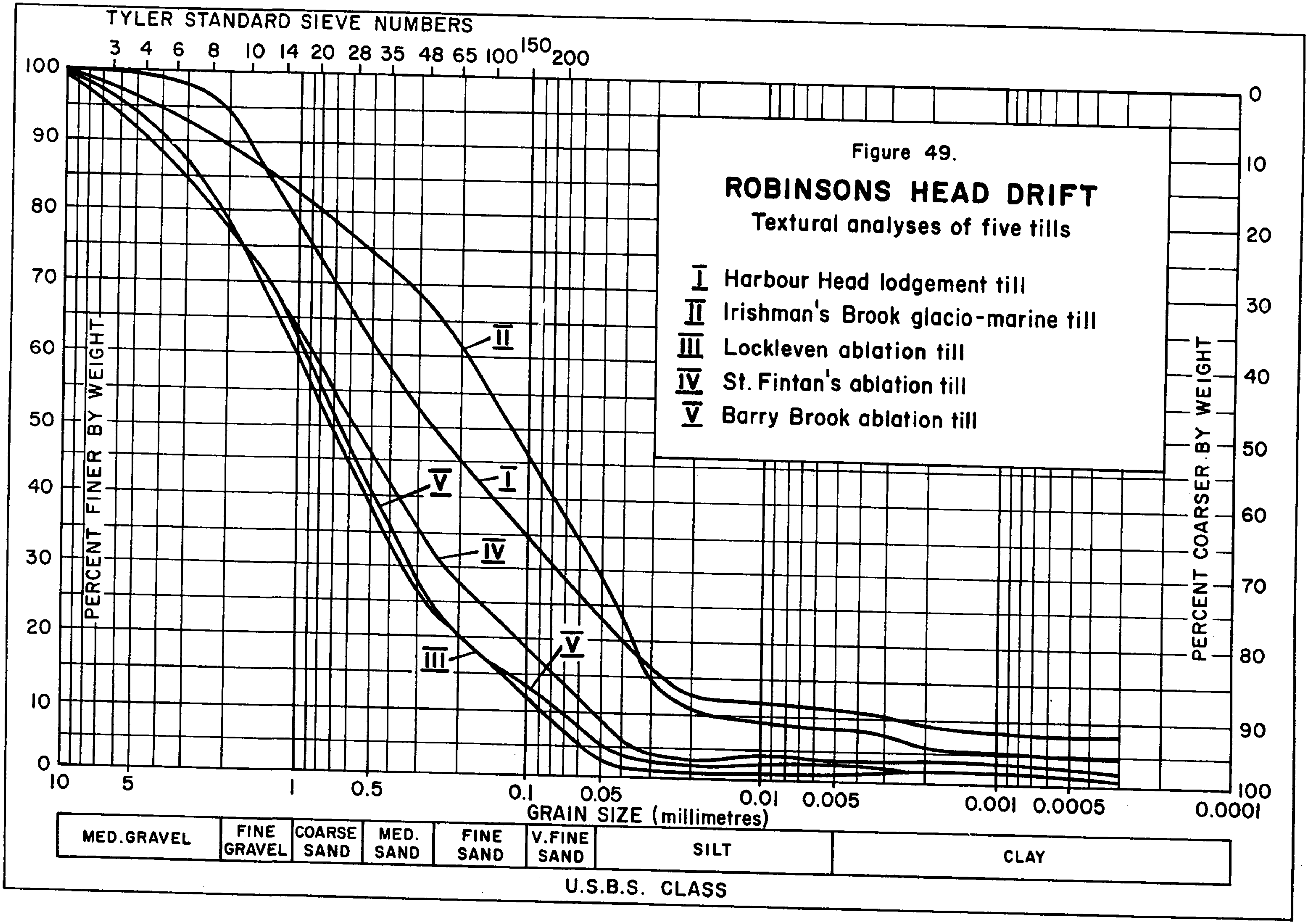


Figure 49.

ROBINSONS HEAD DRIFT

Textural analyses of five tills

I Harbour Head lodgement till
II Irishman's Brook glacio-marine till
III Lockleven ablation till
IV St. Fintan's ablation till
V Barry Brook ablation till

a. The Harbour Head to Bank Head Section (Plate IX)

This part of the St. George's Bay shore shows two major terrain types: the hummocky surface of the frontal lobes of the Robinsons Head Drift end moraine, with lower, irregular morainic terrain between and inland of the lobes, and the level surface of glacio-marine delta fragments. The surficial deposits and landforms of the Harbour Head locality and the stratigraphy of the cliff section are shown in figure 45. In Chapter Five, the nature of marine overlap at Harbour Head was discussed and it was concluded that the till in the upper part of the cliff face was deposited by ice moving across the surface of a glacio-marine delta. The late-glacial sea level had fallen to 88 feet at this time from a higher marine limit stand. River cliff sections in stream valleys in this area show marine deposits up to about 200 yards inland. Ice may have retreated further inland than this but it is unlikely to have been far more than a quarter of a mile. This is indicated by the nature of the junction of the Robinsons Head Drift and the marine terrace at this and many other localities around St. George's Bay. The junction is only a quarter of a mile inland and is low and indistinct. It shows several shallow, flat-floored meltwater channels leading from the moraine onto the terrace surface. No features indicative of overriding of the terrace surface, or of wave-cutting of the moraine are seen. Only on the flanks of the higher end moraine ridge in the cliff face is the zone of boulder concentration beneath the 90-foot terrace indicative of wave action. The ice appears, therefore, to have surged forward only locally and to have merely halted in its retreat between the salient end moraine lobes.

At Robinsons Head coastal cliff sections show marine deposits extending to at least 116 feet above HHWL, and possibly to 140 feet. Above the marine deposits, well-stratified kame gravels dip in many different directions as a reflection of ice-contact depositional conditions. An example of one cliff section, and the plan form of the end moraine are shown in figure 42. While marine deposits occur to at least 116 feet there, glacial deposits of the end moraine are found below this elevation. It is evident, again, that the local surge with which this part of the moraine is associated moved across a sea floor. At Jeffreys, south of Robinsons Head, a terrace-moraine junction similar to that at Harbour Head, stands at 84 feet. In these two sites, therefore, the evidence indicates that the late-glacial sea had been higher than the level of the terrace which abuts against the moraine, and that the reactivation of the ice sheet produced local surges, with marginal stillstands in intervening areas, when the sea had been lowered to the 84-foot level.

At Bank Head, stratigraphic relations between marine and younger glacial deposits are similar to those discussed above. In the headland itself, marine deposits occur only to 80 feet, above which they give way to a complex of kame gravels and till pockets which builds the upper faces of cliffs to more than 150 feet. End moraine summits inland of these cliffs rise to 400 feet above sea level, but it is likely that Carboniferous bedrock rises higher there than to the north and south. A prominent ice-contact face, with large kettle lakes on its proximal side attests to the lobate form of the ice margin here.

South of Bank Head, coastal cliffs display a sequence of deltaic sediments which rises to the cliff tops at 125-130 feet. The surface of this delta rises very gently inland for two miles to an indistinct junction

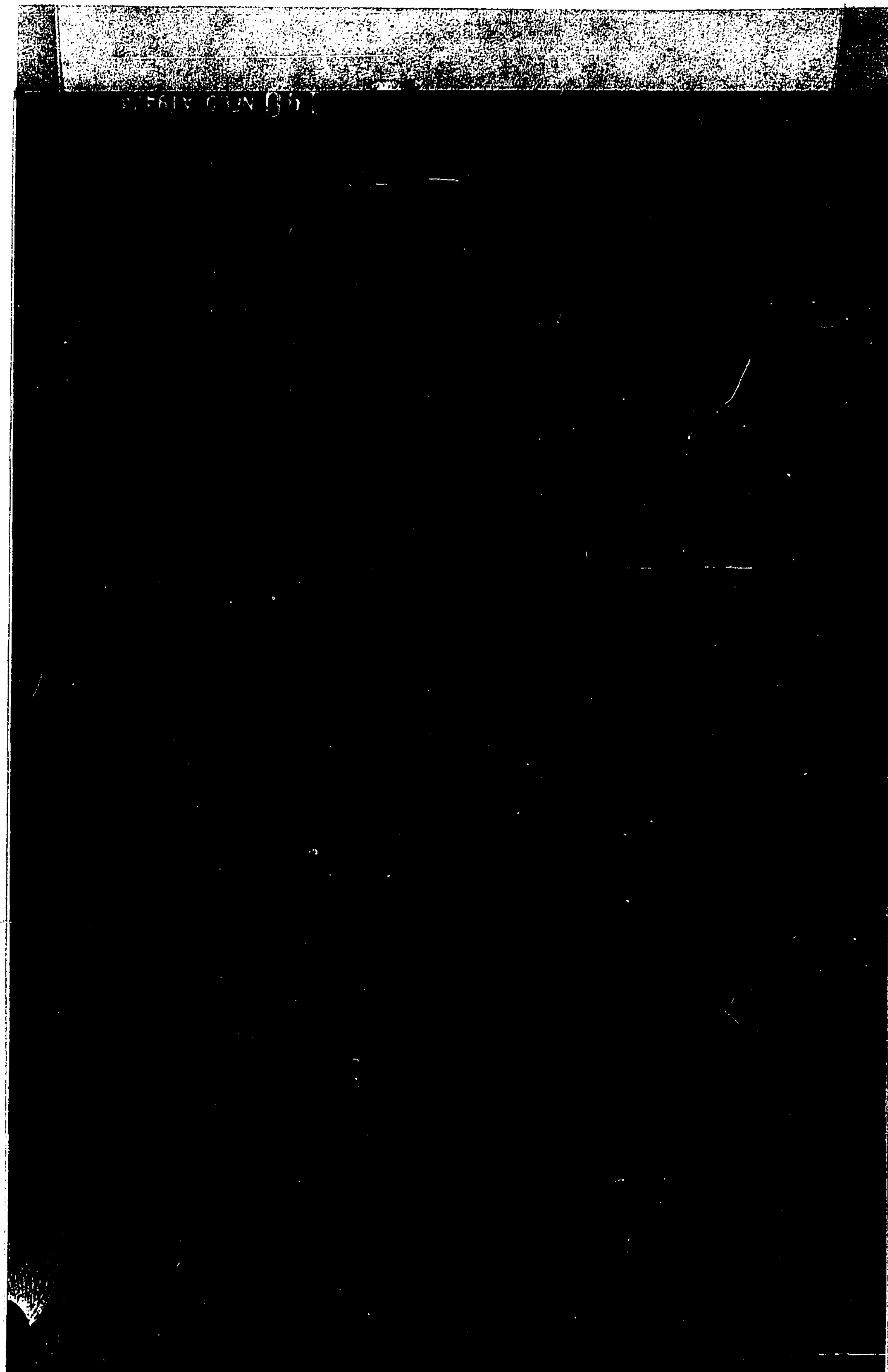
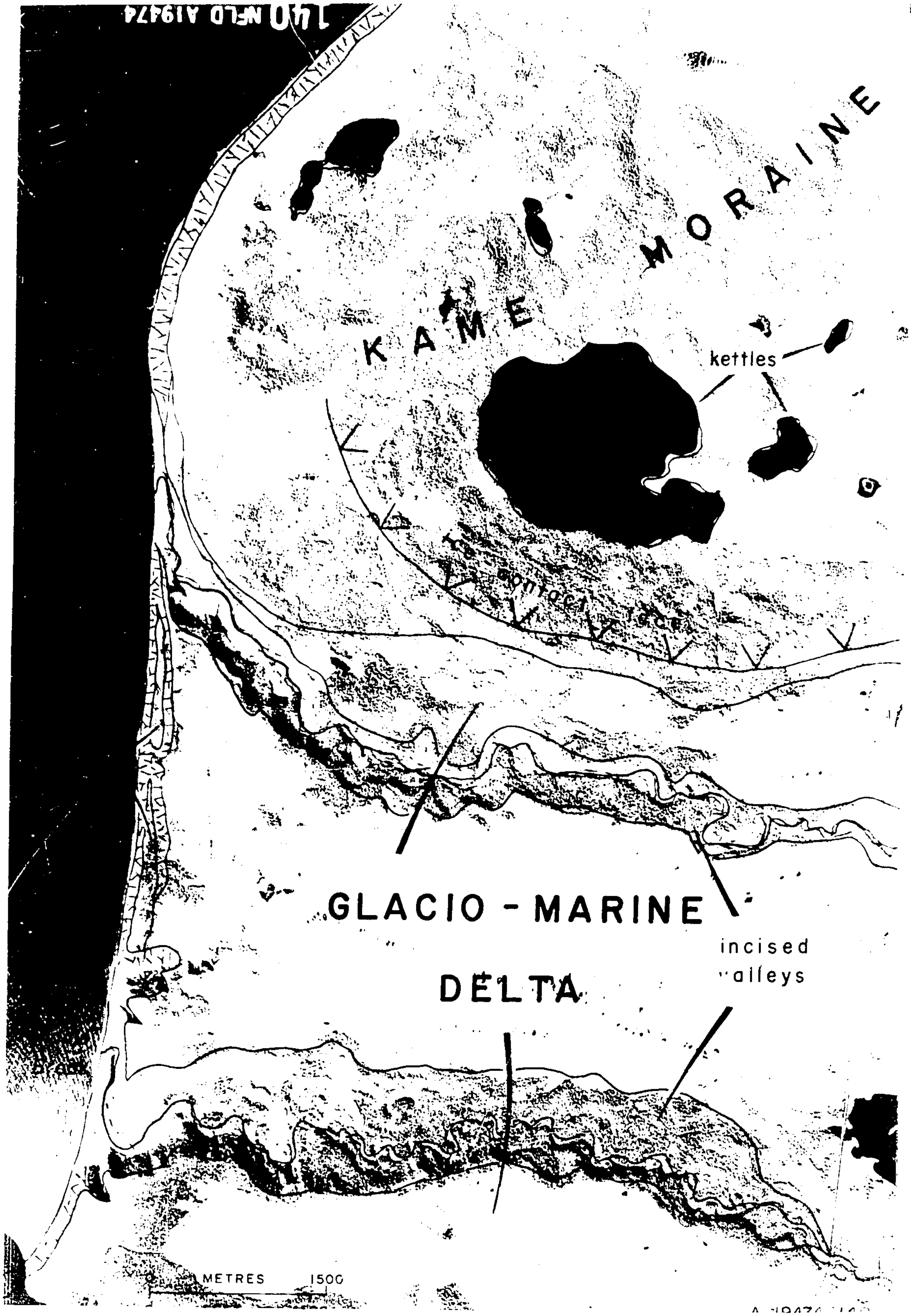


Figure 50: Junction of end moraine and raised delta, Bank Head.



KAME MORAINE
kettles
ice contact face
GLACIO - MARINE
DELTA
incised
valleys
0 METRES 1500

with the low terrain of the Robinson Head Drift at 160-170 feet. The northern limits of this delta surface, however, are marked by an abrupt junction with the lobe of end moraine which builds Bank Head (figure 50). While it has been shown that, at Robinsons Head and Harbour Head, the formation of the end moraine coincided with the 85-foot sea level stand, the delta-moraine junction at 160-170 feet south of Bank Head would seem to negate this. The altitude of the junction may, however, be explained as the result of a stillstand in ice recession on a delta surface which had been built at the marine limit and which remained intact, apparently unaffected by lower terracing. This terracing may, however, have occurred further seawards, but cliff recession has destroyed any evidence of it.

The valleys of Journois, Middle, and Barry Brooks are cut into the 130-160-foot delta surface and they have river terrace fragments preserved on their sides. Those on the north side of Middle Brook were levelled in this work. The back of the bluffs of these terraces, along only one survey line, lies at 24, 57, 94, and 117 feet. The 94-foot terrace may represent a valley floor graded to the 84-foot sea level which was responsible for the terrace flats seen at Harbour Head and Robinsons Head. This level is also seen north of Bank Head, south of the valley of Flat Bay Brook, so that its absence at Bank Head is more likely due to cliff recession than to factors which prevented its registration.

b. The Harrys River Lowland (Plate X)

The wide, sweeping meanders of Harrys River cut through the end moraine of the Robinsons Head Drift at Black Duck, near the road bridge across the river. In this area the morainic topography is markedly irregular, exhibiting hummocky kame-and-kettle form. A stagnant ice-front is indicated. Further indications of this are seen in the esker system which flanks the Indian Head Range, passing westwards into the over-steepened trough of Long Gull Pond.

Southwest of Black Duck, Harrys River flows in a seasonally flooded, shallow channel, with many bars and high-water channels. Above the modern channel are three terrace levels, noticeable only on the west side. The extent of the terraces can be seen in figure 51. Spot heights obtained by levelling allow simple longitudinal profiles of the upper two terraces to be drawn. The middle one slopes south-southwest at 1:180 and the upper in the same direction at 1:420. The proximal part of the upper terrace, at the east end of Long Gull Pond, is at an elevation of 120-130 feet. It is covered with low ridges of glaciofluvial deposits which are sandy "aprons" spread around the distal parts of eskers. This relation suggests a synchroneity of esker and terrace formation, but the eskers could also be interpreted as later deposits spread over the terrace surface.

The middle terrace is at 95 feet above sea level at its proximal end, southeast of Long Gull Pond, and is separated from the upper terrace by a prominent bluff 50 feet high. Gravel pits cut into this bluff show that the deposits of the upper terrace are coarse, well-sorted and well-rounded gravels, mostly of cobble size. They have sub-horizontal bedding that is suggestive of torrential deposition by "shooting flow" (figure 52). Similar deposits are seen in exposures in the second terrace bluff. The bluff separating these two terraces has a broadly arcuate form, and small bog rivulets which cross it show a right-angled change of direction from perpendicular to, to parallel to, the bluff. This suggests that the

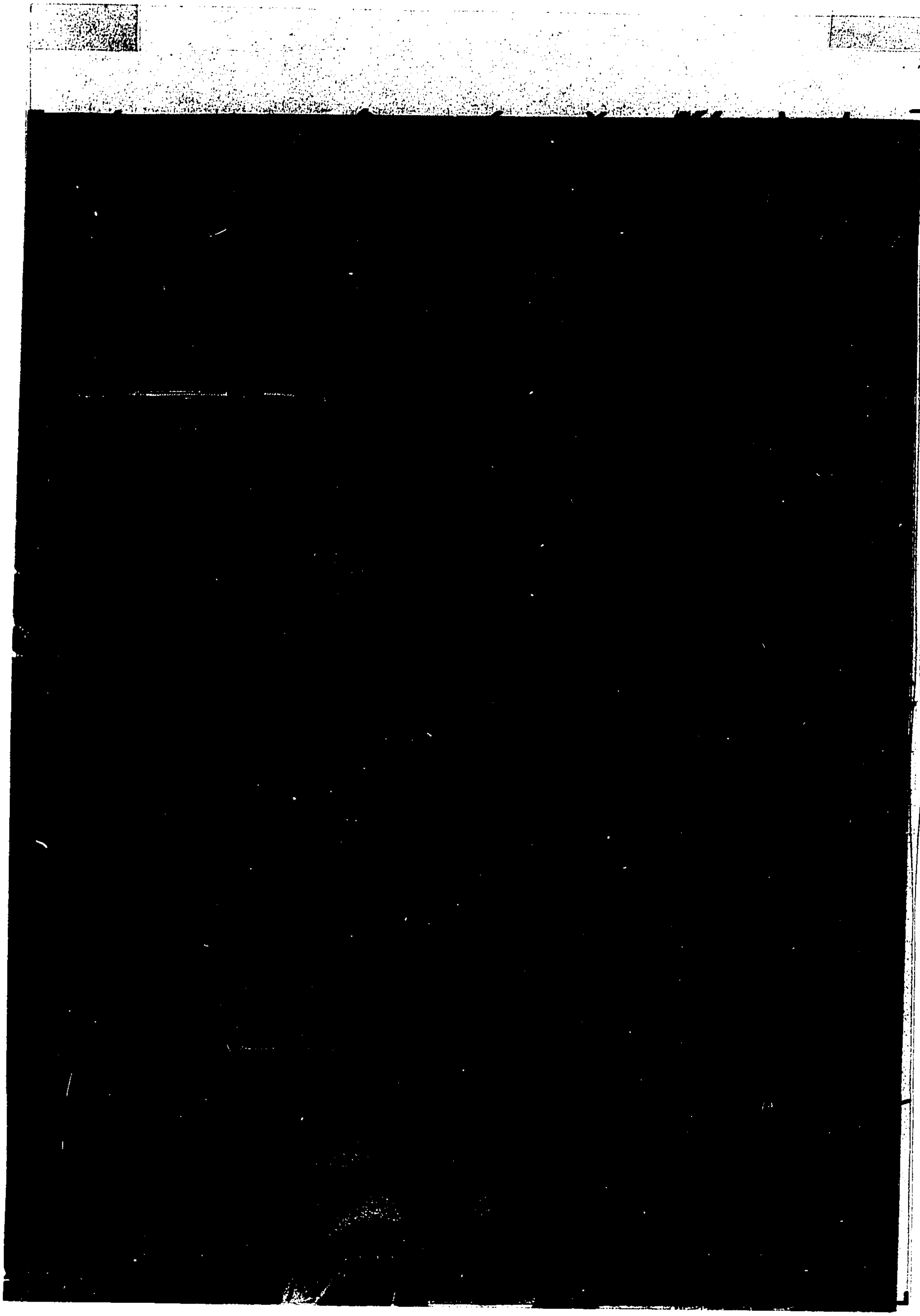


Figure 51: Terraces on the west side of the Harrys River lowland.


St.
River
St.
River
156-169

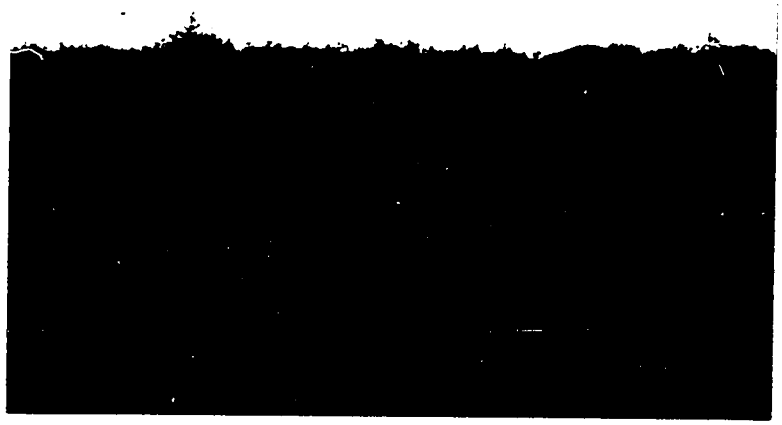


Figure 52: Coarse gravels of upper terrace, west side of Harrys River Lowland.

Figure 52: Coarse gravels of upper terrace, west side of Harrys River Lowland.

bluffs were cut by meandering estuarine channels and that the streams developed on the flats which emerged as sea level was gradually lowered.

The deposition of the upper terrace gravels could not have begun until ice in the Harrys River lowland had wasted back as far as the southern limit of glaciofluvial deposits in the Long Gull Pond area (Plate X). These deposits meet the proximal margin of the terrace at 120-130 feet. The higher of these two figures is the elevation of the peat surface above the surface of the terrace gravels and the lower figure is the elevation of the gravel surface on the shore of Long Gull Pond. The actual elevation of the junction of glacial deposits and terrace is, therefore, closer to 120 feet, and this may be taken to be the upper marine limit in the lowland.

The middle terrace, therefore, represents a regressional still-stand in the lowering of sea level since deglaciation, and may be the correlative of the 85 to 90-foot regressional level along the eastern shore of St. George's Bay. The proximal margin of this terrace merges with the glaciofluvial deposits near Black Duck, in the same way as that of the upper terrace. Since these margins are only 20 feet apart vertically, it is possible that the lower terrace was also built in contact with decaying glacier ice near that settlement. The rate of sea level lowering has been estimated in Chapter Seven at nine feet per century, so that only 200 years would separate the formation of these terraces.

On the east side of the lower part of the Harrys River valley, there is no such convenient association of terraces and end moraine which enables the sequence of events to be reconstructed. From the edge of the modern channel, minor terrace flats, which are probably not related to sea level changes, are bounded by a prominent bluff which marks the distal edge of a very wide, gently sloping area upon which no features can be detected which would betray its origin. This level area itself is bounded to the east by a steeper slope which rises on to a level, boggy surface underlain by St. George's River Drift (Plate X). Along this break in slope, small streams draining northwestwards change direction to southwest. This suggests that the break in slope is a marine bluff and that streams which cross it were extended southwestwards across an emerging marine level as the sea was lowered in that direction. The elevation of this bluff could not be levelled in this study, but air photographs reveal that the foot of it rises gently from about 100 feet to near 200 feet in a northeasterly direction. The bluff must be marine in origin up to 120 feet, since that is the elevation of the upper marine limit on the west side of the lowland. But it cannot be marine at higher elevations. The level area probably represents a terrace, most of which is of subaerial origin, deposited by an earlier Harrys River graded to the marine limit elevation of 120 feet. No dissection of the terrace occurred on the scale of that which affected the terraces across the river, since the river underwent a continuous lateral swing away from the bluff described above, to a course closer to the western side of the lowland. It is also possible that Trout Brook, a left-bank tributary of Harrys River was responsible for depositing this wide terrace. This stream appears to have once continued across the terrace in a southwesterly direction, but today turns 90 degrees to the right to join Harrys River below the terrace, while a small bog stream continues across the latter as the likely beheaded remnant of its lower reaches.

The wide expanse of St. George's River Drift north of the tidewater inlet of that name was not inundated during marine overlap since it stood above the marine limit. Further, it was not subsequently covered by Robinsons Head Drift since, upon the decay of ice in this lowland which permitted marine overlap to occur, the ice was not reactivated to surge in lobate projections to seaward as it was in other places along the St. George's Bay shoreline. The most active ice in the Harrys River lowland remained along the flank of the Indian Head Range. During its decay, the complex esker system was deposited. The southernmost member of this system passes into Long Gull Pond. At the stand of the ice margin marked by the end moraine at Black Duck, there was, therefore, a lobe of ice in the Long Gull Pond trough which fed meltwaters and debris westwards into the late-glacial sea in the Harmon field area. This latter area is discussed below.

c. The Harmon Field to Port au Port Section (Plate XI)

i. The Harmon Field area. Stephenville Pond is a shallow, freshwater lagoon, once cut off from the sea by a magnificent series of parallel boulder beaches, but now opened to it by a man-made cut. From the edge of the lagoon a steep, scrub-covered, sea cliff rises onto a wide, boggy expanse of raised delta standing at 95 feet above HHWL. The delta surface is strewn with many shallow ponds. These are kettle holes which indicate that delta deposition took place in close association with stagnant ice. The proximal margin of this delta surface drops steeply to the floors of irregular, deeper kettles which are interspersed with irregular linear and conical hummocks of coarse, ill-sorted drift (figure 53). This kame-and-kettle area marks the outer limit of Robinsons Head Drift, and it is

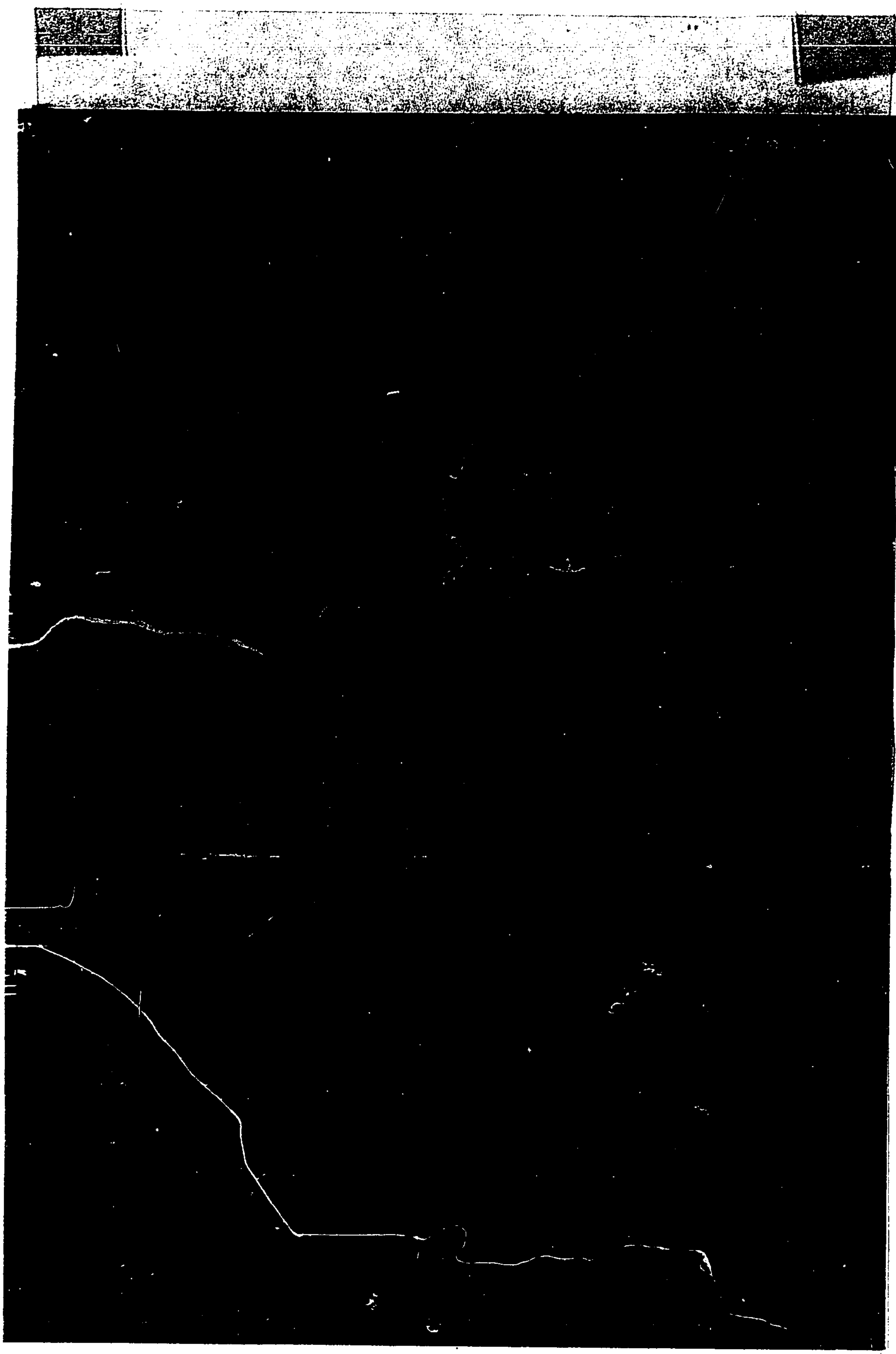


Figure 53: The Harmon Field area: pitted delta surface and kame moraine.

ground moraine
kame and kettle
Flood Plain
pitted delta
moraine
Stephenville
Pond
0 METRES 1500
0 FEET 4500

contiguous to the east with the esker system in Long Gull Pond, and to the west with end moraine topography that trends southwestwards to the shore of St. George's Bay near Kippens.

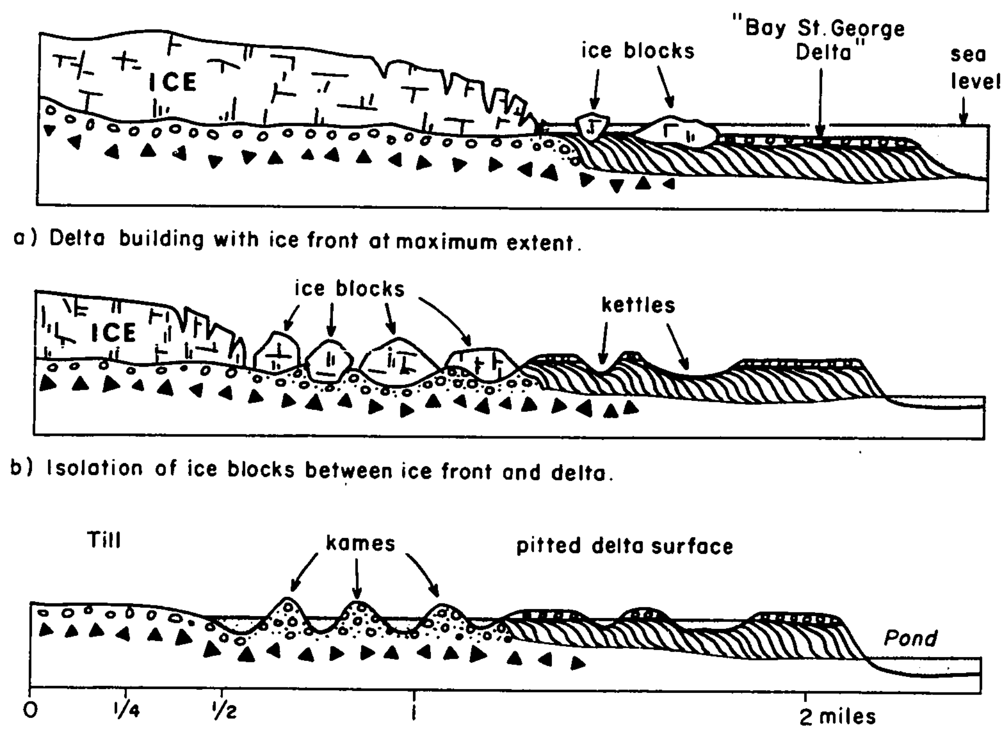
The moraine-delta relationship indicates that ice did not clear the embayment west of the Indian Head Range until sea level had fallen to 95 feet. At that time the margin of the decaying ice mass was washed by the sea at 95 feet and into this sea the Harmon Field delta terrace was built. This is, therefore, contemporaneous with the formation of the middle terrace level in the Harrys River lowland. The deglaciation of that lowland, therefore, occurred at a slightly earlier time than in the Harmon Field area. This can be attributed to the activity of the ice lobe which remained in the trough of Long Gull Pond and ice which had flowed from the high areas of the Indian Head Range and the Lewis Hills. The probable sequence of events is shown in figure 54.

ii. <u>The Harmon Field to Port au Port Section (Plate XI)</u>. The end moraine of the Robinsons Head Drift trends northeast to southwest behind the town of Stephenville and meets the present shoreline at the western limits of the town. Between there and the west side of the mouth of Romaines Brook, the coastal cliffs have been cut into the distal margin of the moraine, so that it is evident that this lobation extended up to one mile beyond the present shoreline. The cliff top rises and falls through the surface of kame hills and kettle depressions, providing optimal conditions for the investigation of the deposits in the cliff faces (figure 55).

Between Stephenville and Port au Port, the exposures in the cliff faces display three kinds of sedimentary and topographic sequences. The

Figure 54.

Evolution of landforms in the Harmon Field area.



c) Present landscape with kames and kettle lakes behind pitted delta surface.

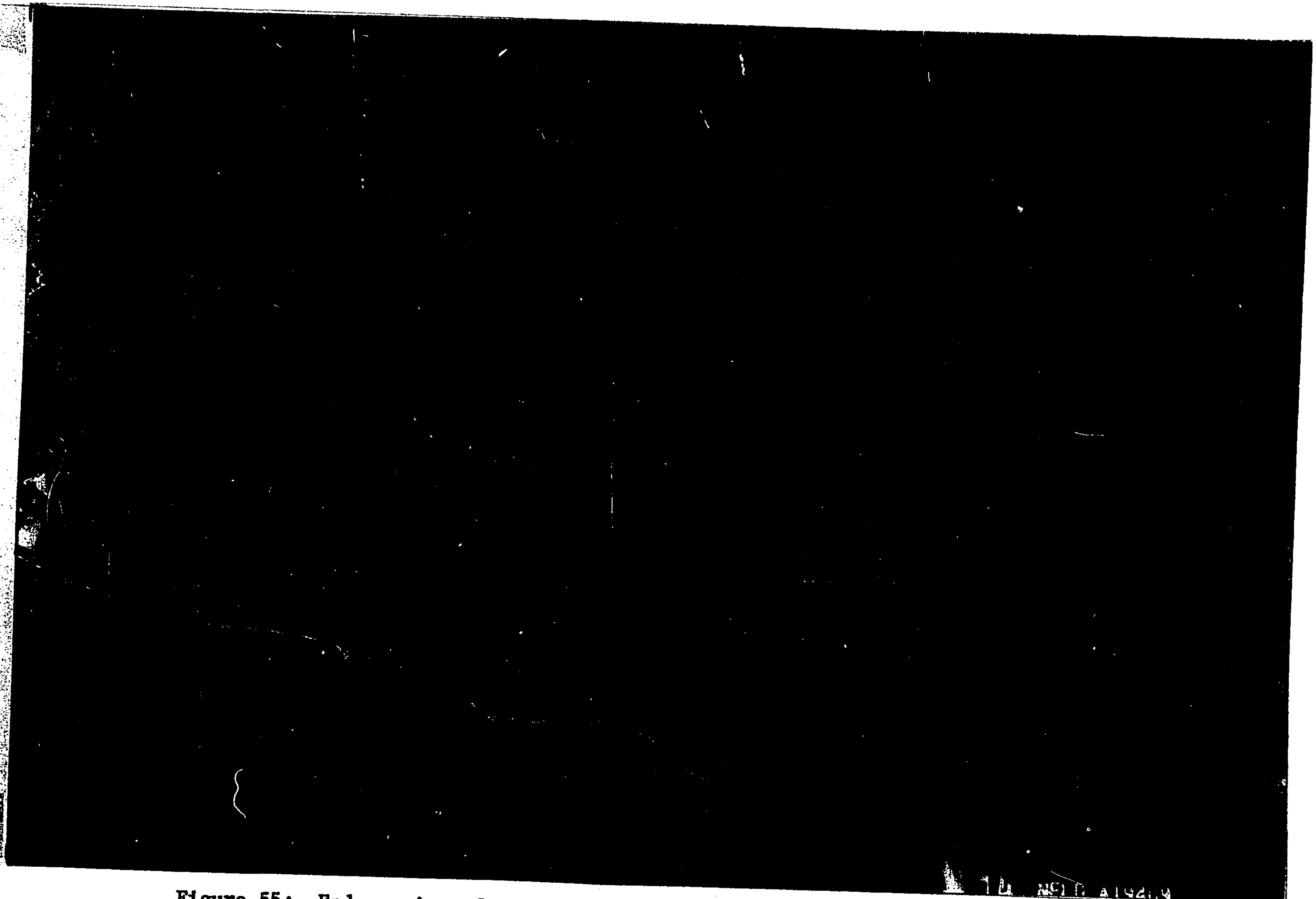
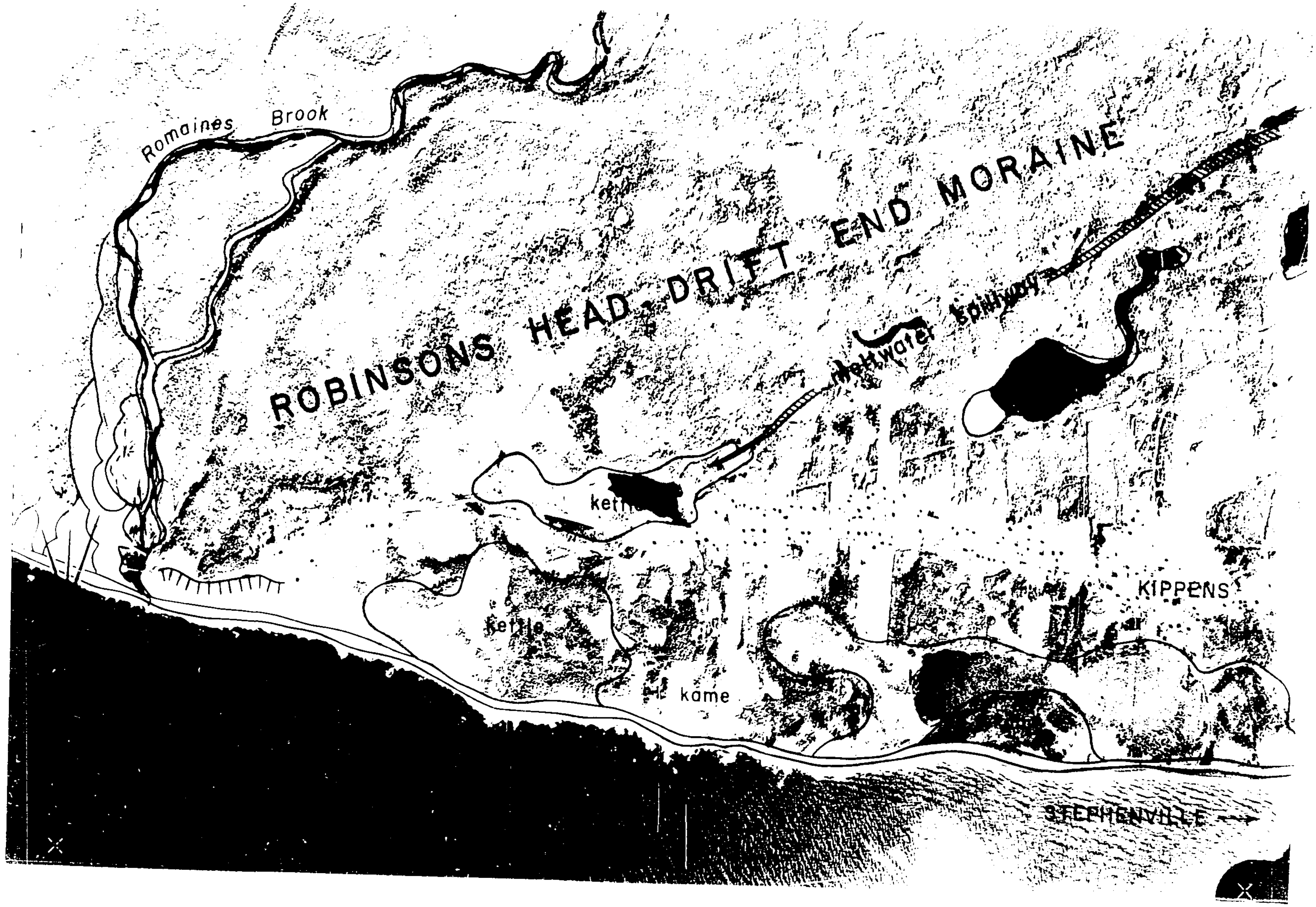


Figure 55: End moraine of Robinsons Head Drift, west of Stephenville.


Romaines Brook
ROBINSONS HEAD DRIFT END MORAINE
Meltwater spillway
kettle
kettle
kame
KIPPENS
STEPHENVILLE

most common is that where kame and kettles have been cut by the cliffs. There, the cliffs show fine-grained marine deposits, with shells, draped over the undulating surface of the St. George's River Drift (figure 56). Delta foreset beds overlie these to a maximum elevation of 80 feet. Above these, the regular bedding and uniform texture of the sands give way to coarser, unevenly stratified gravels, with many cobbles and boulders, which are the kame materials. The summits of the kames are at 90-125 feet above sea level and they slope steeply down to the floors of breached kettle depressions which stand at 25-30 feet. In these kettles, no coarse deltaic sediments or glaciofluvium are found. The top of the finer-grained marine deposits lies within a few feet of the kettle floor. Apparently, delta foreset deposition was prevented in these kettle depressions, but fine-grained marine sediments were laid down. Isolated ice blocks must have been present along the present shoreline during delta, and subsequent kame, formation adjacent to them. In order that these blocks could prevent delta deposition they must have been stranded beyond the wasting ice front during the initial deglaciation of the shoreline. However, ice could not have retreated very far inland since the ice blocks were still present during the subsequent formation of kames at the Robinsons Head Drift stage. A closely-spaced sequence of events is indicated in which it is not as meaningful, as in localities on the east shore of St. George's Bay, to refer to two separate glacial events. It appears that, before blocks left by the wastage of St. George's River Drift ice could be melted, ice was again present along this shoreline and kames were being formed. A short marine phase intervened, in which basal fine-grained sediments and delta foresets were deposited.

Figure 56: Coastal cliffs west of Stephenville, showing basal marine silts draped over pockets of ice-contact stratified debris half-way up cliff-face. Delta foresets above and till at base.

Figure 56: Coastal cliffs west of Stephenville, showing basal marine silts draped over pockets of ice-contact stratified debris half-way up cliff-face. Delta foresets above and till at base.

Figure 56: Coastal cliffs west of Stephenville, showing basal marine silts draped over pockets of ice-contact stratified debris half-way up cliff-face. Delta foresets above and till at base.

This conclusion is difficult to harmonize with evidence from the valley of Romaines Brook. There, marine incursion extended almost a mile inland of the present shoreline. In a river cliff opposite the western side of the large bar which divides the channel of Romaines Brook, silts and clays with marine shells occur below delta foreset sands. These, in turn, are overlain by till of the Robinsons Head Drift. A greater inland retreat of ice may have been possible here due to the topographic "low" of the valley which is cut deeply into bedrock. Certainly, great differences in depositional conditions over short distances are called for.

Another of the three depositional sequences seen in coastal cliffs in this section is that which occurs west of Romaines Brook. A 125-foot summit west of the mouth of that stream is a kame hill marking the furthest extent to the west of the Robinsons Head Drift. Inland of the summit the margin of the drift fades indistinctly against the lower slopes of Table Mountain (Plate XI). Between this kame hill and the isthmus at Port au Port, coastal cliffs rise to a level surface at 100-115 feet above sea level. The deposits exposed in the cliff faces comprise St. George's River Drift till at the base, with delta bottomset and foreset beds above, continuing to the cliff top. They are, then, all marine in origin, and the sequence of events here has been discussed in Chapter Five.

The third depositional-topographic form seen in this area is restricted to a small stretch of shoreline east of Romaines Brook. There, the delta foreset beds have been truncated by a terrace which stands at 63 feet above sea level. The terrace bluff is cut into the end moraine of the Robinsons Head Drift and, with the river terraces so well developed on the west side of Romaines Brook, it is the only evidence of regressional sea level stands along this shore. It is significant that these terraces are localized here since it was postulated above that the valley of this stream appears to have allowed the ingress of the late-glacial sea further inland than in other localities. Drainage of that embayment during the postglacial period of rapid uplift would have left evidence of regressional stillstands within it. No doubt these stillstands were in evidence against the distal margin of the Robinsons Head Drift end moraine all along this shoreline, but cliff recession has destroyed their remains.

In summary, a well-developed marine terrace occurs at the western and eastern extremities of this area of Robinsons Head Drift. That at Port au Port, at 113 feet above sea level, was formed at about 13,400 years B.P. The lower level of the terrace at Harmon Field indicates a slightly later formation. Marine overlap occurred up to one mile inland up the valley of Romaines Brook, but the glacier did not retreat far inland of the present shoreline in the middle section of the area. Further, in that area, in the subsequent Robinsons Head Drift stage, the ice surged beyond the shoreline in a lobe. At Harmon Field it merely halted in its retreat since it was less well-nourished.

This discussion of the Robinsons Head Drift event has been restricted to the evidence provided in the St. George's Bay area. This is partly because it is in that area that the stratigraphic and topographic relations of the drift are clearly seen. Other end moraines are seen north of St. George's Bay, in the valley of Fox Island River and south of Serpentine River, but the relations of these to the Robinsons Head Drift cannot be discussed until the topic of land and sea level changes has been dealt with. There is no contiguity between them and the St. George's

Bay moraine and no stratigraphic sections are found. Indirect evidence of their position in the scheme of events will be provided in Chapter Seven.

CHAPTER SEVEN

Postglacial Changes of Level in Southwestern Newfoundland

1. INTRODUCTION

This section is intended to summarize the discussion of the late-glacial period of marine overlap in southwestern Newfoundland, presented in Chapter Five, in order to provide a link between that chapter and this, in which the topic of land- and sea-level changes will be reviewed in its entirety.

Radiocarbon dates on late-glacial overlap of Atlantic waters in southeastern Canada show that the event occupied the relatively short interval between 13,700 and 12,600 years B.P. (figure 57). This rapid flooding was assisted by three factors. Firstly, the ice front was being calved into relatively warm sea waters. Secondly, because of glacio-isostatic depression, sea level, which had been rising eustatically since about 17,000 years B.P., was high up on the outer edges of the ice and could, therefore, melt large amounts of ice. Thirdly, this calving process was especially rapid along the two arms and main trunk of the deep Laurentian Channel. Glaciers were, no doubt, floating at their margins above the floor of the Channel while remaining grounded on shallow sea bottoms which are now gently shelving offshore zones. Evidence of the rapidity of this process is found in the date of 12,940 ± 180 years B.P. (G.S.C.-89) (Dyck and Fyles, 1963), for marine overlap on western Anticosti Island, and that of 12,410 ± 170 years B.P. (G.S.C.-101) (Dyck and Fyles, 1963), close in time to the same event in northern Prince Edward Island.

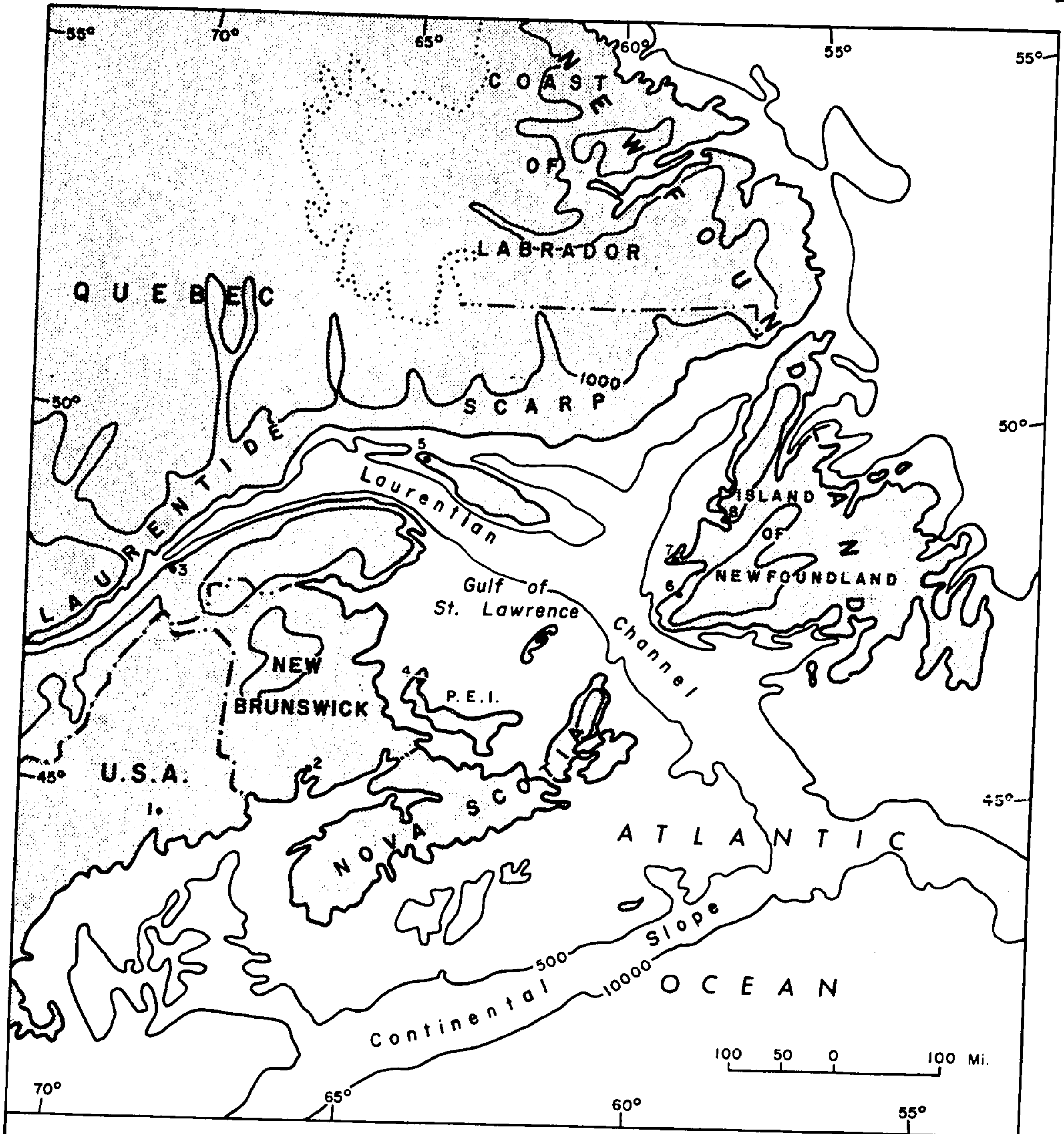


Figure 57.

Radiocarbon dates on the start of marine overlap in the Canadian Atlantic Provinces.

•1 Radiocarbon-dated site with age X 10³ years B.P. 12,800 ± 450 (W-1011)	•5 12,940 ± 180 (G.S.C.-89)
•2 13,200 ± 200 (G.S.C.-965)	•6 13,500 ± 210 (G.S.C.-1200)
•3 12,720 ± 170 (G.S.C.-102)	•7 13,700 ± 230 (G.S.C.-1074)
•4 12,670 ± 340 (G.S.C.-160)	•8 12,600 ± 170 (G.S.C.-868)
	—500— Isobath in feet.

The latter site is 200 miles closer to Cabot Strait than Anticosti Island, but is fringed by a gently shelving offshore zone, whereas Anticosti lies close to the deep water of the Laurentian Channel.

Radiocarbon dates from southwestern Newfoundland show that the onset of marine overlap occupied the time interval between 13,700 ± 230 years (G.S.C.-1074) and 12,600 ± 170 years B.P. (G.S.C.-868). The earlier dates are from sites closer to Cabot Strait and at greater distances from ice outflow centres. The stratigraphy of marine deposits at radiocarbon-dated sites in the area indicates that the sea flooded the newly deglaciated littoral zone to depths ranging from about 90 feet at Robinsons Head to more than 160 feet at Cox's Cove. These figures are obtained from the altitudinal differences between the marine limit and the surface of glacial deposits beneath the marine sediments. Where the marine limit has not been obliterated by younger glacial deposits, variation in its elevation is due both to the variation in the date of its formation and to differential isostatic warping.

2. THE CAUSES OF VARIATION IN THE ELEVATION OF MARINE LIMIT FEATURES

The map showing the elevation of raised marine features in southwestern Newfoundland, Plate VIII, shows that features generally rise in elevation along the north-northeast trend of the coast. Differential isostatic uplift over the area can be immediately suspected as contributing to this. However, the demonstrated differences in the age of marine limit features further suggests that this has added a complication. Before the relative role of each of these factors can be evaluated, it is necessary to determine the form of uplift curves from sites where sufficient data are available.

a. Isostatic Uplift

i. General considerations. The derivation of an uplift curve relies upon an accurate determination of the fall of sea level in the postglacial period, marked by dated, regressional marine features. Each feature has been uplifted by an amount equal to its present elevation above sea level plus the amount by which sea level itself has risen eustatically since the formation of the feature. A trace of absolute uplift through postglacial time is obtained by adding to the elevation of each dated horizon a figure for eustatic sea level rise since that date.

In southwestern Newfoundland, the paucity of well-defined, dateable, regressional marine features below the marine limit prevents the empirical derivation of an uplift curve for any one site, or even a composite curve from a number of sites. The regressional features will be discussed in the following section, but it can be said here that they are well marked in only a few places and, in those places where shell fragments were obtained from them, it was impossible to relate the shell-bearing horizon to a stand of sea level.

In this chapter an indirect approach will be made to the determination of the postglacial changes in sea level. Hypothetical uplift curves are derived for two sites for which an elevation and an age for the marine limit are available. The method used is that recently developed by Andrews (1968a). These curves are used in conjunction with an eustatic sea level curve plotted by Shepard and Curray (1967). The picture of the changes of sea level derived here is intended only as a model with which available field data can be compared and from which further studies can proceed.

The eustatic changes in the level of the sea since the last glacial maximum have been a subject of a debate which began with the workers in the Mediterranean region (Depéret, 1926; Blanc, 1937). Flint (1957) was able to present only a discouraging picture of the state of knowledge regarding glacio-eustatic sea level changes thirteen years ago. Since that time, however, the number of studies has increased many-fold, although the debate over the significance of results continues. Radiocarbon dating has been of obvious importance in establishing temporal equivalence between curves from widely separated areas.

Today, the number of curves of eustatic sea level rise since the last glacial maximum is embarrassingly large for the investigator wishing to select one for studies of glacio-isostasy. Until recently, it was thought to be suitable to choose a curve which showed the general trend in several areas of differing stability, or one which was accurately reconstructed for an area thought to be stable. The continental shelf of eastern North America, south of the limits of Pleistocene glaciation, has until recently been thought of as relatively stable for the purpose of obtaining an eustatic sea level curve (Curray, 1965). Many studies in the last few years have been directed to obtaining curves from this region (Emery and Garrison, 1967; Milliman and Emery, 1968; Scholl and Stuiver, 1967; Redfield, 1967) and Shepard and Curray (1967) have attempted to give a unified picture.

Bloom (1967) attributed differences in the postglacial submergence histories of sites along the eastern seaboard of the United States to differential subsidence of the offshore zone beneath the weight of an eustatically rising sea. According to Bloom, and others who have recognized this "hydroisostatic" effect (Wellman, 1964; Higgins, 1969; Grant, 1970a), no offshore zone can be regarded as having remained stable during the post-maximum submergence of late-Wisconsin time. The difficulty of isolating the "hydroisostatic" effect from that of eustatic sea-level rise, glacio-isostatic uplift, and other tectonic disturbances, poses a whole new set of problems in studies of sea level changes, whether or not an area was previously covered by ice sheets.

In this investigation of glacio-isostatic uplift in southwestern Newfoundland, some careful assumptions have to be made in order to estimate the form of that uplift. Several questions may be asked. How much crustal depression occurred at the glacial maximum? Was this depression in isostatic equilibrium with the ice load? What proportion of the depression (if this was in isostatic equilibrium) has occurred as restrained and postglacial rebound, and how much has yet to occur as residual rebound? Finally, what was the rate of uplift in each interval of time between deglaciation and the present?

The simplest approach to the first question is to view glacio-isostatic depression as the same fraction of former maximum ice thickness as the ratio of the densities of ice and the displaced sub-crustal material. The ice-sheet over western Newfoundland covered 2,500-foot summits at the glacial maximum. At this stage ice was also grounded on the present sea floor where depths today are approximately 400 feet. Further, late-glacial erosion scoured fjords to depths of 800 feet below present sea level in the Bay of Islands. If it is assumed that the shallower rises between fjord basins represent parts of the pre-late-glacial valley floor, ice extended to about 500 feet below present sea level. If the 2,500-foot summits were covered by flowing ice, a thickness of 500 feet can be assumed as a minimum. The thickness of ice over western Newfoundland at the glacial maximum can, therefore, be estimated at about 3,500 feet.

Another approach to the estimation of ice thickness is the use of empirical equations derived for existing ice cap profiles. These were used in connection with the discussion of the vertical extent of glaciation at the Wisconsin maximum, in Chapter Four. There, an equation of the form

$$h = 100\ D^{0.57}$$

where h = elevation of ice surface above base, in metres, at D; and
D = distance from ice margin, in kilometres

described an ice-surface profile parallel to the direction of ice flow, which gave a minimum thickness of 100 metres of ice over the highest summits of western Newfoundland. Those summits are known to have been glaciated at the maximum, and fresh erratics atop them attest to the former presence of actively flowing ice.

Figure 27 shows that Abrahams Cove was overlain by 840 metres of ice. For the fjord site of Cox's Cove, a figure of 1,065 metres is comparable to that of 3,500 feet (1,060 metres) estimated for the high-relief areas from topographic evidence. The comparability cannot be viewed over-confidently since, in the latter case, only a reasoned estimate of 500 feet was made for ice thickness over the highest summits.

Glacio-isostatic depression in equilibrium with an ice load is the same fraction of former ice thickness as the ratio of the specific gravity of glacier ice to that of the displaced sub-crustal material. With a specific gravity of ice equal to 0.9 and that of sub-crustal material equal to 3.3, total depression at Abrahams Cove is given by the following calculation,

$$\frac{840 \times 0.9}{3.3} = \text{ca. } 230 \text{ metres, or } 760 \text{ feet}$$

and depression at Cox's Cove by

$$\frac{1065 \times 0.9}{3.3} = \text{ca. } 290 \text{ metres, or } 960 \text{ feet.}$$

The second question, as to the degree of isostatic equilibrium attained by the crust in response to ice loads as large and as long-lasting as those of the Wisconsin ice sheets, continues to be argued. Many complex geophysical considerations need to be accounted, including the determination of loads, their duration, and crustal properties, but these are outside the scope of this study. Some attempt will be made, however, to justify the assumption made below, that isostatic equilibrium was achieved between the mass of the Newfoundland ice cap and the crust and mantle beneath.

Gutenberg (1941) showed that the time required to reduce the stress resulting from a load on the crust to 1/e of a former value, if the strain remains constant, is of the order of 3×10^3 years. He warned against the confusion of this with the time in which the strain is reduced to 1/e of a former value if the load is removed. This was shown to be 10 times that of the former parameter, and it increases with time. Andrews (1968a) has reached a similar conclusion from an analysis of arctic Canadian postglacial uplift curves. Gutenberg showed that its order of magnitude is 10^4 years. Perhaps this time parameter may be thought of in reverse; that is, a similar order of magnitude is required to establish isostatic equilibrium with an ice cap load. Crittenden (1970) finds that "loads of long wavelength (> 200 km) and of long duration (> 20,000 years) will attain isostatic balance with a high degree of precision" (pp. 727-728).

Recent estimates of the duration of the last glacial sub-stage, in which ice sheets attained their greatest mass and extent (for example, Ericson and Wollin, 1964; Morrison and Frye, 1965; Dreimanis, 1967), agree that from 50,000 to 20,000 years ago a glacial maximum was in effect. Various authors differ on the status of a warmer period which preceded it, and of warm periods within it. Notwithstanding these differences, there is general agreement that glaciers were at their widest extent during the period 50,000-20,000 years B.P. This period probably was sufficiently long for isostatic equilibrium to be achieved between ice masses and sub-crustal layers.

The third question posed above, as to the amount of total isostatic uplift which occurred before deglaciation, and which has occurred since that time, must be approached via the uplift curves themselves, since these give an approximation to postglacial uplift. The final question of the actual form of postglacial uplift will be discussed in the following sub-section.

ii. Models of land and sea-level change. With these foregoing geophysical considerations in mind, the form of postglacial uplift in southwestern Newfoundland can be discussed. Due to the absence of dated (and dateable) marine features below the marine limit, uplift curves will be approximated using the general equation of Andrews (1968a). Andrews' analysis of 21 uplift curves for sites in arctic Canada showed that the form of the curves was described by the equations

$$U' = Ce^{-kt} \qquad (a)$$

where U' is the amount of uplift, in metres, remaining after $\underline{t}$ years, $\underline{C}$ is amount of uplift remaining at $\underline{t} = 0.0$ (i.e., recorded postglacial uplift), $\underline{k}$ is a "decay" constant that varies with time, and $\underline{t}$ is years $\times 10^3$ since deglaciation,

and
$$U = \frac{A(1-i^t)}{1-i} \qquad (b)$$

where $\underline{U}$ is uplift accomplished in $\underline{t}$ years since deglaciation, $\underline{A}$ is a constant for each site which is the percentage of total uplift accomplished in the first 1.0×10^3 years after deglaciation, $\underline{i}$ is a constant (given a value of 0.677 for arctic Canada), and is the value by which a preceding term can be multiplied to give a value for uplift in a time interval of $\underline{t}$ years.

In equation (a) the "decay" constant $\underline{k}$ was obtained "by considering the uplift accomplished in the first 1.0×10^3 years as a percentage of uplift accomplished in 2.0, 3.0×10^3 years, etc." (Andrews, 1968a, p. 44). Values for k in equations describing 21 uplift curves in arctic Canada range from 0.37 to 0.56, and Andrews attributes these departures from predicted values to sources of error in determining marine limit elevations, positions of eustatic sea level at deglaciation, and radiocarbon dating.

In this study a k value of 0.40 was used to arrive at an approximate uplift curve for two sites, Abrahams Cove and Cox's Cove[1]. Equation (b) was not used in this study since the value of i given by Andrews (1968a) may not apply to this area, and the value of A is uncertain.

At the Abrahams Cove site, the marine limit elevation is 140 feet (42.4 metres) and its age is 13,700 ± 170 years B.P. (G.S.C.-1074). Eustatic sea level rise had attained a level of 238 feet (72 metres) below present at that date, so total recorded uplift, C, is 376 feet (114 metres). A value of 0.40 is assigned to k in the formula

$$U' = Ce^{-kt}$$

which can be expressed in logarithms to base 10, as

$$\log_{10}U' = \log_{10}114 - (0.4343 \times 0.40 \times t).$$

For values of t from 1.0 to 13.0 x 10^3 years after deglaciation, the values of U' are given in Table III, page 152.

At the Cox's Cove site, the marine limit elevation is 166 feet (48.5 metres), and its age is 12,600 ± 170 years B.P. (G.S.C.-868). With the eustatic correction, the total postglacial uplift recorded is 366 feet (112 metres). Table IV (page 153) shows the values of uplift remaining in each 1,000-year interval after deglaciation, and the percentage of total uplift taking place in each interval.

The uplift curves plotted from the U' values shown above are

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1. This value was advised by J. T. Andrews (personal communication, 1970).

TABLE III

POSTGLACIAL ISOSTATIC UPLIFT AT ABRAHAMS COVE, ST. GEORGE'S BAY

Years x 10^3 after deglaciation	U' uplift remaining (m)	% total uplift accomplished	% total uplift accomplished (in Andrews, 1968a)[a]
1	76.4	33.0	33.0 ± 4.4
2	51.2	55.0	56.0 ± 5.7
3	34.3	70.0	70.4 ± 5.8
4	23.0	80.0	80.2 ± 5.4
5	15.4	86.5	86.5 ± 4.5
6	10.3	90.0	91.1 ± 3.7
7	6.9	94.0	93.9 ± 2.9
8	4.6	96.0	94.4 ± 3.0
9	3.1	97.0	97.4 ± 1.9
10	2.1	98.0	100
11	1.4	99.0	*
12	0.9	99.3	*
13	0.6	99.5	*

[a]Andrews, 1968a, Table II, p. 42.

*Andrews did not continue his uplift curved beyond 10^4 years after the deglaciation of a site.

TABLE IV

POSTGLACIAL ISOSTATIC UPLIFT AT COX'S COVE, BAY OF ISLANDS

Years x 10^3 after deglaciation	U' uplift remaining (m)	% total uplift accomplished	% total uplift accomplished (in Andrews, 1968a)[a]
1	75.1	32.5	33.0 ± 4.4
2	50.3	55.0	56 ± 5.7
3	33.7	70.0	70.4 ± 5.8
4	22.6	80.0	80.2 ± 5.4
5	16.2	85.0	86.5 ± 4.5
6	10.2	91.0	91.1 ± 3.7
7	6.8	94.0	93.9 ± 2.9
8	4.6	96.0	94.4 ± 3.0
9	3.1	97.0	97.4 ± 1.9
10	2.1	98.0	100
11	1.4	98.7	*
12	0.9	99.2	*

[a]Andrews, 1968a, Table II, p. 42.

*see footnote Table III.

shown in figures 58 and 59. The percentages of uplift achieved in each 1.0×10^3 years closely conform to the average calculated by Andrews (1968a).

On the model of land and sea level change at Abrahams Cove (figure 58), the eustatic sea-level curve is taken from Shepard and Curray (1967) after 16,000 years B.P. The resultant of these two curves is the curve of the position of sea level relative to present through time. A rapid fall of 9 feet (2.7 metres) per century in the first 1,000 years after deglaciation, and a rate only slightly lower for the next 1,000 years, caused sea level to reach its present level at about 11,800 years B.P. The uplift rate first equalled the eustatic rise of sea level at 10,000 years B.P., and a maximum emergence of 33 feet (10 metres) lasted for the next 800 years. Following that, eustatic sea-level rise was greater than uplift and submergence occurred back up to present sea level. According to this model, sea level has been higher than 16.5 feet (5 metres) below its present level since 7,600 years B.P.

These models of postglacial sea level change can be used to evaluate the contribution of (i) the variability in the ages of marine limit features and (ii) differential isostatic warping, on the elevation of these features. They also provide a basis for a discussion of restrained rebound (sub-section v).

iii. <u>The effect of variable age on the elevation of marine limit features.</u> The radiocarbon dates from the sites discussed in Chapter Five indicate a general synchroneity of marine overlap of the shores of St. George's and Port au Port Bays. If the standard error is subtracted from the oldest

Figure 58.

A MODEL OF POSTGLACIAL LAND AND SEA-LEVEL CHANGE AT ABRAHAMS COVE, ST. GEORGE'S BAY

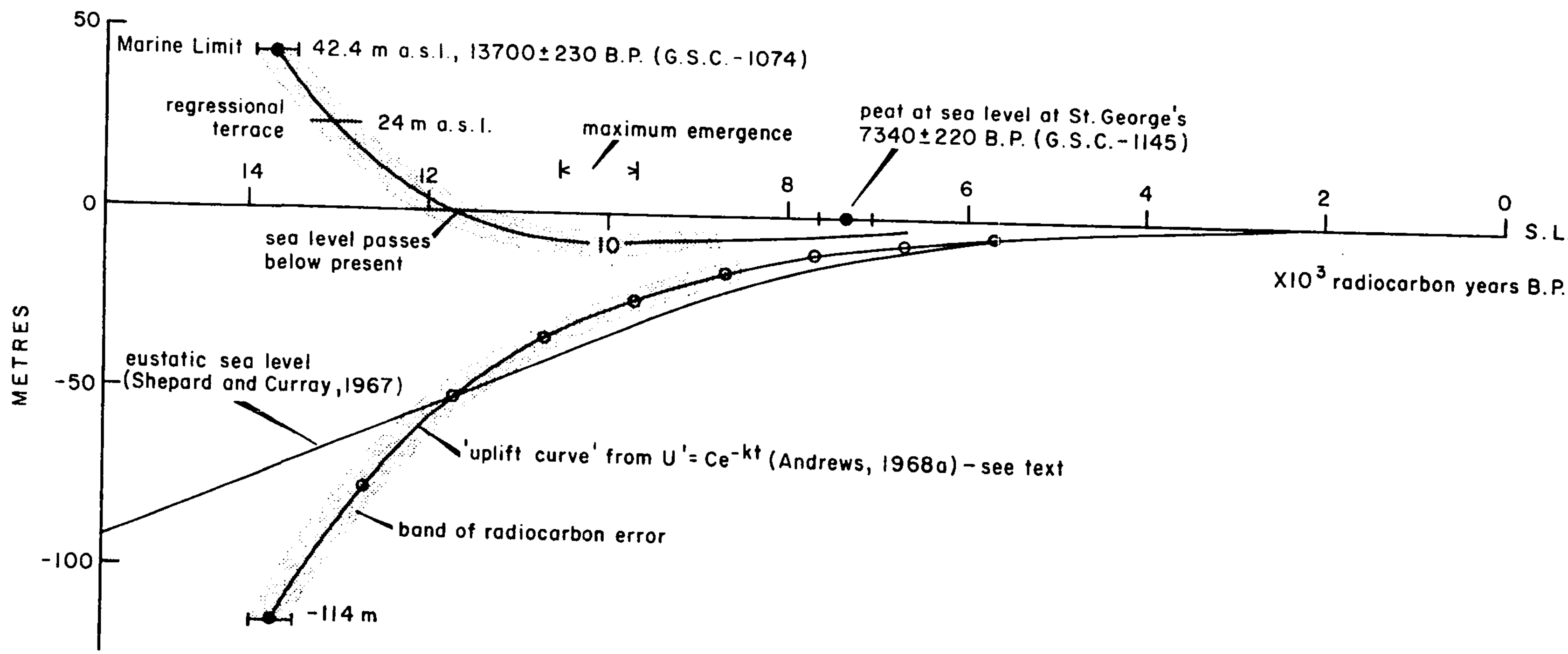
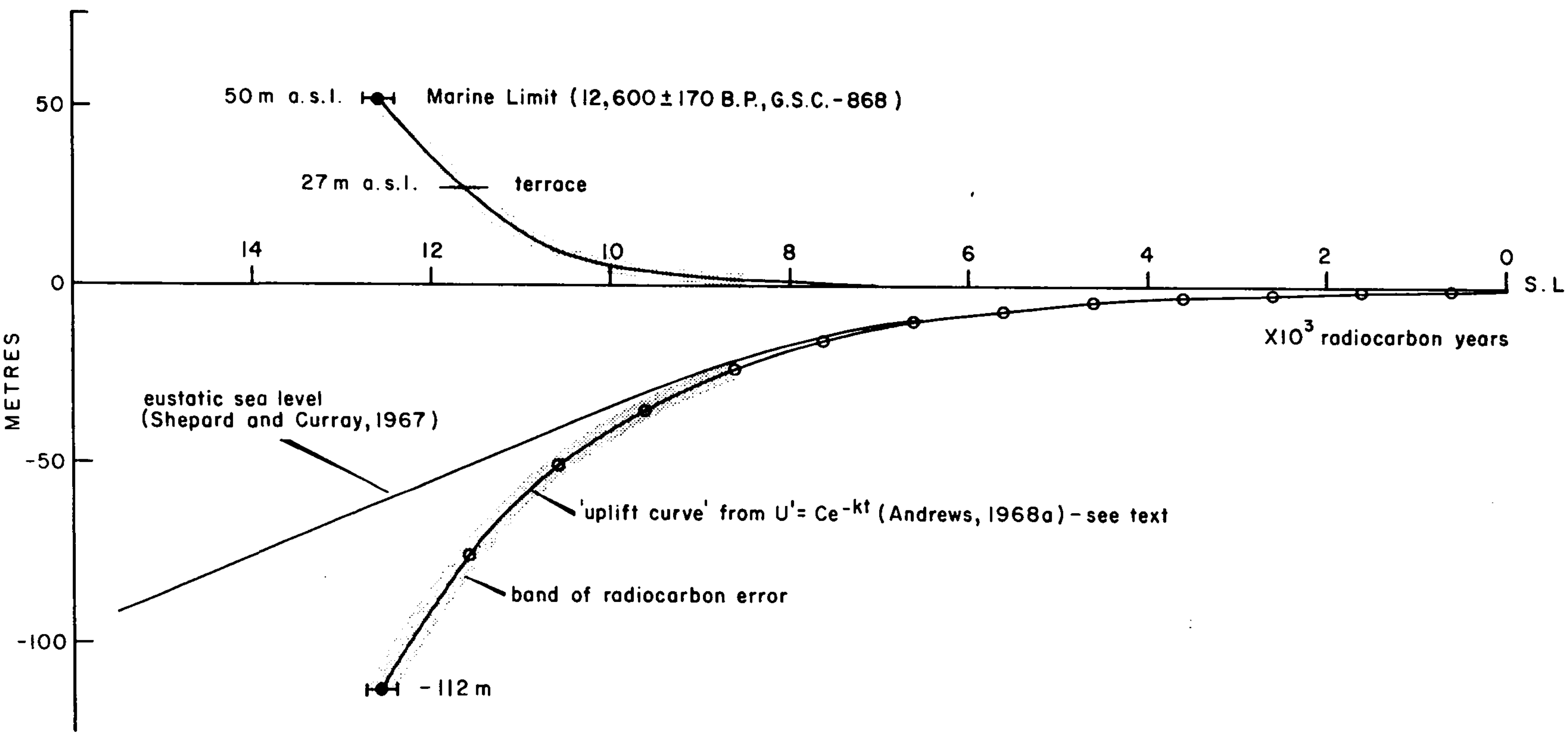


Figure 59.

A MODEL OF POSTGLACIAL LAND AND SEA-LEVEL CHANGE AT COX'S COVE, BAY OF ISLANDS



date of 13,700 ± 230 (G.S.C.-1074, from Abrahams Cove) and added to the youngest date of 13,200 ± 220 (G.S.C.-937, from Tea Cove) there is a difference of only 50 years. The range in the central value of each date (see Table II), might suggest that there is no reason to suspect that variations in marine limit elevations are due to the different dates of their registration. However, the rate of isostatic uplift in this time interval should be taken into account.

Andrews (1969) has pointed to the necessity of considering the standard error of radiocarbon dates in comparing the elevations of late-glacial marine features. He asserts the necessity of doubling the 1σ error quoted by the dating laboratory, so that there is a 95.45% chance of the radiocarbon age being included within the given limits. If a 2σ error is applied to the dates from St. George's Bay they can all be considered to be equal.

While I recognize the uncertainties introduced by considering only the central values of the dates, I feel that this is justifiable in the case of the present study because, firstly, of the scanty nature of the evidence which is available from western Newfoundland, and, secondly, because of the importance of arriving at even a tentative reconstruction of dateable events in that area.

The events with which this study is concerned occurred between approximately 14,000 and 12,500 years ago, and the standard errors of the radiocarbon dates range from 170 years to 290 years. In contrast, in the areas from which Andrews drew the above conclusion, deglaciation occurred 7,000 to 8,000 years ago, and standard errors on radiocarbon dates range from 110 to 175 years. Obviously, a great degree of accuracy is unattainable at present in the area under discussion here, and the tentative nature of the conclusions must be stressed. To some extent their validity can be judged by their internal consistency and by that displayed with respect to supplementary data to be discussed later in this chapter.

The isostatic uplift curve for Abrahams Curve shows a rate of uplift of 3.63 metres (12.0 feet) per century between 13,700 and 12,700 years B.P. With a eustatic sea level rise in the same interval of about 1 metre (3.3 feet) per century, sea level would be falling at 2.63 metres (8.7 feet) per century. The 500-year range of ages for marine limit features in St. George's and Port au Port Bays could alone produce a range of marine limit elevations of 5 x 2.63 or 18.15 metres (43.5 feet). It is noteworthy that the elevation of the dated marine limit at these sites, from at least 100 feet at Tea Cove to 140 feet at Abrahams Cove, is approximately the same as the range which could be due to their different ages.

The date of 12,600 ± 170 years B.P. (G.S.C.-868) for the marine limit at Cox's Cove, Bay of Islands, shows that deglaciation of the fjord was delayed by up to 1,100 years after the ice first cleared the gently-shelving shoreline of the salient Port au Port Peninsula. It has been argued above that a delta terrace at the head of Humber Arm, Bay of Islands, is the same age as that at Cox's Cove, so that the entire bay was ice-free by 12,600 years ago. The upper marine limit in the bay is at 160 feet (48.5 metres), 20 feet (6.1 metres) higher than the highest limit at the St. George's Bay area. If the rate of sea level fall between 13,700 and 12,700 years B.P., calculated above, is applied to the 1,100-year interval between the Cox's Cove and Abrahams Cove dates on the marine limit, 96 feet (29.1 metres) of emergence (i.e., uplift less eustatic sea level rise) occurred at Abrahams Cove between deglaciation there and the same event at Cox's Cove. Expressed another way: if the date of the marine limit was 13,700 years B.P. at both sites, the marine limit at Cox's Cove would now stand at 256 feet (74.5 metres), rather than 160 feet (48.5 metres). This reasoning does not take into account the possible greater rate of isostatic uplift at Cox's Cove in that time interval, which would have occurred if, as seems reasonable, that site lay closer to a centre of ice loading at the glacial maximum. The hypothetical figure calculated above can be interpreted as a minimum. The implications of these differences in the age of the marine limit for the reconstruction of regional isobase trends are discussed in subsection iv of this chapter.

iv. Differential isostatic warping. In the most general sense the elevation of marine limit features shows a rise towards the north-northeast along the southwestern Newfoundland littoral. Features near 140 feet in St. George's Bay, 160 feet in the Bay of Islands, and at 195 feet at Trout River can be used to illustrate this trend (Plate VIII). If the effects of the younger age of marine limits in the Bay of Islands and at Trout River are accounted for, the difference in absolute amounts of uplift experienced between south and north is even more marked. Apparently, differential isostatic uplift has occurred since deglaciation and it shows a strong component along the general north-northeast trend of the coastline.

Difficulties arise in attempting to construct isobases of equal uplift on the marine limit in this area. Obviously, the greatest obstacle is the variability of marine limit ages. But, even where radiocarbon ages are comparable, such as at Abrahams Cove and Robinsons Head, or at Port au Port and Highlands, the elevations are so similar over wide areas that the drawing of one isobase with an exclusive trend is impossible. For example, the marine limit at Abrahams Cove is at 140 feet, at Robinsons Head it is near 140 feet, south of Bank Head it is near 145 feet. An isobase could be drawn between Abrahams Cove and either Robinsons Head or Bank Head, or between the latter two sites. Obviously, the problem cannot be surmounted using the ages and actual elevations of the marine limits.

A tentative solution to the problem can be offered here. I am aware of its shortcomings, but feel that the method to be discussed gives an indication of at least the direction of isostatic warping, if not the real values on each isobase. It is the direction which is important in this area, since the conclusion of Flint (1940), that Labrador ice had invaded western Newfoundland, was based upon the trend of isobases which he reconstructed on upper marine features. His conclusion cannot be harmonized with the evidence of ice-flow directions at the Wisconsin maximum.

The method used here is based upon the rate of the fall of sea level from the marine limit at Abrahams Cove, St. George's Bay. This rate is calculated by subtracting from the isostatic uplift rate (from Andrews formula) the rate of eustatic sea level rise. A rate of 8.7 feet per century results. Abrahams Cove was chosen as it is the site from which the earliest date of marine overlap in the area has been obtained. Figure 60 shows the measured and estimated elevations of selected marine limits in the study area and, where available, their radiocarbon ages. The hypothetical elevations of marine limits were then calculated for all the sites except Abrahams Cove, on the assumption that they were deglaciated at the same time as that site, 13,700 years B.P. The number of centuries of difference in the central values of the radiocarbon age at Abrahams Cove and other sites was multiplied by 8.7. Three sites were then chosen as the apices of a triangle which enabled a line to be drawn joining two sites with equal elevation on this hypothetical "marine limit" plane. The resultant line was taken to be orthogonal to the regional isostatic warping direction, and parallel "isobases" were drawn through sites, using the hypothetical marine limit elevations (figure 60). The lines which pass through Humbermouth-Cox's Cove and Trout River give a gradient to this warped plane of two feet per mile towards the northeast. Continuation of this gradient into the St. George's Bay area shows how close are the isobase values on the hypothetical marine limit to actual marine limits at sites where deglaciation was approximately simultaneous with that at Abrahams Cove.

This method of determining the regional trend of isostatic warping may be checked against another which also arises out of the uplift curves. From curves drawn for Abrahams Cove, Cox's Cove, and Trout River, the elevation of sea level relative to present may be determined at these sites at any particular date. If, for example, the date of 10,000 years B.P. is chosen, sea level stood at -7 metres at Abrahams Cove, 8 metres at Cox's Cove, and 11 metres at Trout River. The triangle method shows that isostatic tilt since that date has approximately the same trend as that demonstrated to have occurred since 13,700 years B.P. The trend of isobases on the 10,000-year plane is a few degrees north of that on the 13,700-year plane. This might be due to arithmetical error arising out of

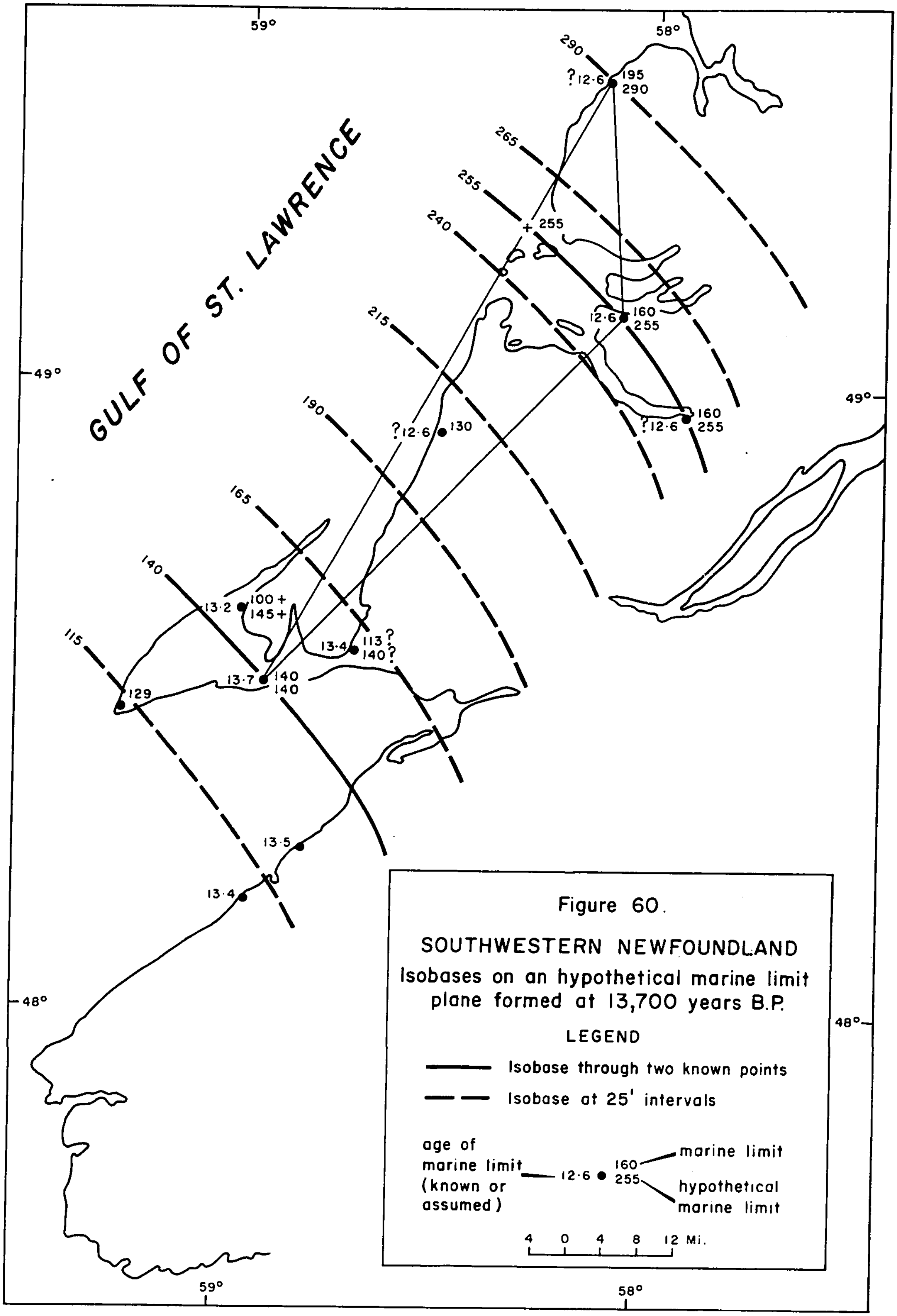


Figure 60.

SOUTHWESTERN NEWFOUNDLAND

Isobases on an hypothetical marine limit plane formed at 13,700 years B.P.

the use of two different methods. However, such a change in the warping trend would be expected if remanent ice masses over the western interior of Newfoundland exerted a greater influence on the pattern of recovery than during the earlier period when thick ice had just begun to clear the western coastal area.

The direction of isostatic warping found by these methods is approximately at right angles to that reconstructed by Flint (1940). The simplest explanation of such a trend is to call for a thickening of Wisconsin ice towards the northeast over western Newfoundland. This is not impossible to comprehend, but it conflicts with the ice-flow directions known for the area: a glacier whose surface sloped up towards the northeast would not flow towards the west at its maximum development.

It was shown in Chapter Four that the western part of the island of Newfoundland was glaciated by an ice cap flowing westward from highland centres of outflow over plateaus that presently lie between 2,000 and 2,600 feet in elevation. It would be expected that such an ice cap had a considerable effect on crustal depression at the glacial maximum, and that, during its decline, local remnant ice masses would have had their own individual effects on restrained rebound rates.

The probable dimensions of the ice cap that covered southwestern Newfoundland which are inferred in this study add to the suspicion that it would have had a noticeable effect on crustal warping in the study area. Nevertheless, if it is agreed for the sake of argument (since, it must be admitted, there is no indisputable evidence in favour) that the Laurentian Channel accommodated much of the active flow of ice from Labrador and directed it towards Cabot Strait, the possibility of a component of isostatic depression towards the Channel cannot be overlooked.

Broecker (1966) has developed a model to relate ice cap thickness, rate of ice retreat, and isostatic deformation of proglacial lake shorelines. One of the implications of his model, which fits observed facts tolerably well, is that "the strength of the crust is sufficiently small to prevent the lateral influence of a continental ice sheet from extending more than a few tens of kilometres beyond its margins" (p. 4777).

On the other hand, Walcott (1970), using the same uplift data from the warped Lake Algonquin shoreline, arrived at a figure of 180 ± 30 kilometres for the outer limits of depression. Differences in the initial assumptions made by these authors may account for the divergence of opinion.

The longitudinal axis of the northeast arm of the Laurentian Channel lies 85 kilometres from the mouth of the Bay of Islands, western Newfoundland, so that according to Walcott one can expect a component of warping due to ice filling, and probably over-filling, the Channel, which is approximately 1,200 metres deep. The trend of isostatic warping in southwestern Newfoundland, tentatively established in figure 60, may represent a combination of warping components, one up towards the axis of the Laurentian Channel, and the other up towards the axis of the Long Range Mountains. The question of whether the Newfoundland ice cap had a greater effect on the warping of a sea level plane below the marine limit is an interesting one for future research.

These views were anticipated both by Daly (1921) and by Flint (1940), although both drew an oversimplified and incorrect picture of the form of postglacial uplift over the island.

The foregoing solution to the problem of determining the regional trend of isostatic warping in this area must be considered tentative by virtue of the fact that the crucial variable, the rate of sea level fall in the early postglacial interval, was obtained using an hypothetical uplift curve. Some degree of confidence may be placed in it, however, for two reasons. Firstly, the uplift curve itself cannot vary significantly in form from that given by Andrews' equation since the only source of error is in the proportionality or "decay" constant, k. The value chosen for the two curves constructed here lies in the lower part of the range shown in Andrews' 21 equations for sites in arctic Canada, and is a value chosen in consultation with Andrews (personal communication, 1970). Secondly, while the form of the uplift curve may depart somewhat from that shown, all the computed values of marine limit elevations on the synchronous plane formed at 13,700 years B.P. would change in the same direction by proportioned amounts, so that the trend of hypothetical "isobases" would remain the same.

v. Restrained rebound. Andrews (1968b) has emphasized the three-phase nature of the isostatic rebound process. The first phase occurs while ice is still present over a site, as "restrained rebound." Recorded "postglacial rebound" is marked by the sum of the elevation of a raised feature above sea level and the eustatic sea level rise since the feature's formation. "Residual rebound" is that part of total rebound which has yet to occur.

The only way in which the rate or duration of restrained rebound can be estimated is through the comparison of marine limit elevations at two sites where deglaciation occurred at different times. If no differential isostatic warping has occurred between them, a rate of restrained rebound can be calculated thus

$$U_{rr} = \frac{H_1 - H_2}{A_1 - A_2} + E$$

where U_{rr} = rate of restrained rebound

H_1, H_2 = elevations of marine limits at sites 1 and 2

A_1, A_2 = ages of marine limits at sites 1 and 2

E = rate of eustatic sea level rise between A_1 and A_2.

Dates from Abrahams Cove and Cox's Cove can be used to estimate the rate of restrained rebound in southwestern Newfoundland. These two sites, however, have been isostatically uplifted by different absolute amounts (rather than merely showing different amounts of postglacial uplift due to the difference in the deglaciation dates). In the tentative reconstruction of the elevations which selected marine features would have at present if they had been deglaciated at the same time as Abrahams Cove, i.e., 13,700 years B.P., the resultant hypothetical marine limit plane slopes up towards the northeast. There is a difference in the hypothetical marine limit elevations at Abrahams Cove and Cox's Cove of 105 feet (32 metres). This is the real amount of differential uplift which has taken place between the two sites since 13,700 years B.P.

If this figure is subtracted from the actual elevation of the marine limit at Cox's Cove, the northerly and most uplifted of the two sites, a figure of 55 feet (16.6 metres) is obtained. This is an approximation to the marine limit elevation at Cox's Cove if no differential uplift had occurred between there and Abrahams Cove. The marine limit at Abrahams

Cove now stands at 140 feet (42.4 metres), so that 140 minus 55, or 85, feet (26 metres) of restrained rebound took place at Cox's Cove between the deglaciation there and the deglaciation of Abrahams Cove. That time interval is approximately 1,000 years. Therefore, restrained rebound occurred at a rate given by

$$\frac{(85}{(10} + E)$$

where E is the rate of eustatic sea level rise in the above interval. This last was approximately 3.3 feet (1 metre) per century, which gives a rate of 11.8 feet (3.60 metres) per century for restrained rebound between 13.7 and 12.7 x 10^3 years B.P.

Glacio-isostatic depression at Abrahams Cove has been estimated at 230 metres, and at Cox's Cove at 290 metres. Recorded postglacial uplift at Abrahams Cove is 114 metres. Assuming that this uplift has <u>just</u> been completed at the present day, 230 - 114 = 116 metres of uplift took place before deglaciation. At the rate of 3.60 metres per century, restrained rebound began 3,200 years before deglaciation, or about 16,900 years B.P. At Cox's Cove, recorded postglacial uplift is 112 metres, therefore, 290 - 112 = 178 metres of uplift occurred before deglaciation. At the above rate of restrained rebound, uplift began 4,900 years before deglaciation or 17,500 years B.P. The calculations assume that uplift has only just been completed, and that the calculated rate of restrained rebound was constant from the start. The first assumption seems to be justified from the evidence of rising tide levels in southeastern Canada (Dohler and Ku, 1970), and submerged peats and a submerged forest bed (Grant, personal communication, 1970) in the study area. Also, Andrews

(1970) shows uplift in the area to be less than 0.1 metre per century at present. The second assumption is, admittedly, less firmly based and is the subject of continuing debate.

3. REGRESSIONAL MARINE FEATURES AND THEIR INTERPRETATION

The discussion of regressional marine features has been postponed to this section since they provide very little evidence from which an uplift curve could be derived. These features can now be viewed within the framework provided by the foregoing tentative reconstruction of postglacial land and sea level changes.

The form and distribution of marine features below the marine limit is shown in Plate VIII. Usually only one feature at a site is at all prominent. The subdued form or complete absence of other features can be explained by the rapid rate of sea level lowering after deglaciation. Marine landforms recording a stillstand in the position of sea level were developed rapidly during a total available period of between 2,000 years (shown in the Abrahams Cove model) and 3,000 years (shown in the Cox's Cove model).

a. Terraces

i. The Robinsons Head Drift moraine-terrace junction and the age of end moraines. In St. George's Bay a prominent regressional stillstand is marked by the terrace surface formed penecontemporaneously with the end moraine of the Robinsons Head Drift. This terrace is at close to 85 feet along the east shore of St. George's Bay and fragments of terraces are at 84 feet at the mouth of Romaines Brook, 80 feet at Abrahams Cove, and 92 feet at Cape St. George. At Harmon Field, the surface of a delta built into the sea in contact with ice of the Robinsons Head Drift "maximum," lies at 95 feet. While the difference in the elevation of these terraces is real and requires explanation, it cannot be called real when transferred to the curve of falling sea level at Abrahams Cove. The error attached to the radiocarbon date on the marine limit there necessitates the drawing of a "band" of isostatic uplift and, therefore, sea level lowering. The 85-95 feet values for the terrace fragments are enclosed by the band of error. The curve shows, however, that sea level had fallen to 85 to 95 feet above present approximately 13.0 x 10^3 years B.P.

A radiocarbon date of 12,600 ± 170 years B.P. (G.S.C.-868) for the deglaciation of Middle Arm, Bay of Islands, and, by analogy, for the fjord-head stand of ice in that bay and Bonne Bay, indicates that this stand was not correlative with the Robinsons Head Drift event around St. George's Bay. At the date tentatively assigned to the latter event in this bay the fjords of the Bay of Islands and Bonne Bay were still ice-bound and any reactivation of the ice margin, if it affected ice in these bays, would have produced features now drowned by sea level rise.

The interpretation made of the date of 13,200 ± 220 years B.P. (G.S.C.-937) for the influx of marine waters over Tea Cove, West Bay, Port au Port suggested that Port au Port Bay was still ice-bound at the time the Robinsons Head Drift event was affecting the shores of St. George's Bay. This is further indicated by the absence of raised marine features around Port au Port Bay comparable in elevation to the "85-foot" terrace formed against the end moraine of that drift. An end moraine in the lower valley of Fox Island River (Plate VIII) cannot, therefore, relate to the Robinsons

Head Drift event, since the ice margin lay around the shores of Port au Port Bay: it must be younger.

The delay in the deglaciation of Port au Port Bay, past the time when the shores of St. George's Bay became ice-free may tentatively be explained as follows:

As late-glacial marine waters extended farther and farther northeastwards up the arm of the Laurentian Channel off western Newfoundland, the margin of ice flowing from the coastal mountains over the Port au Port Peninsula would have been calved into a northeast-southwest orientation above the edge of the Channel. Immediately prior to the marine overlap of the south shore of the Port au Port Peninsula, at 13,700 years B.P., ice flow had been oriented generally north to south over the peninsula. With the calving of the ice margin to a southwest-to-northeast orientation off the west coast of the peninsula and north of it, ice flow would be induced to return to its direction at the glacial maximum, to the west. This flow off the Lewis Hills might have been responsible for severing a low lobe of ice in Port au Port Bay from its source of nourishment. The sequence of ice margin changes is shown in figure 61.

The end moraine in the lower Fox Island River valley has been assigned to an age younger than 13,000 years B.P. Similar morainic features along the northern foot slope of the Lewis Hills can be assigned to this position. These features and associated marine landforms are depicted in Plate XII and an aerial photo of the features is shown in figure 62. If the fjord valleys of the Bay of Islands were not flooded by marine waters until 12,600 years B.P., it is reasonable to suppose that the area south of Serpentine River and north of the Lewis Hills was not ice-free

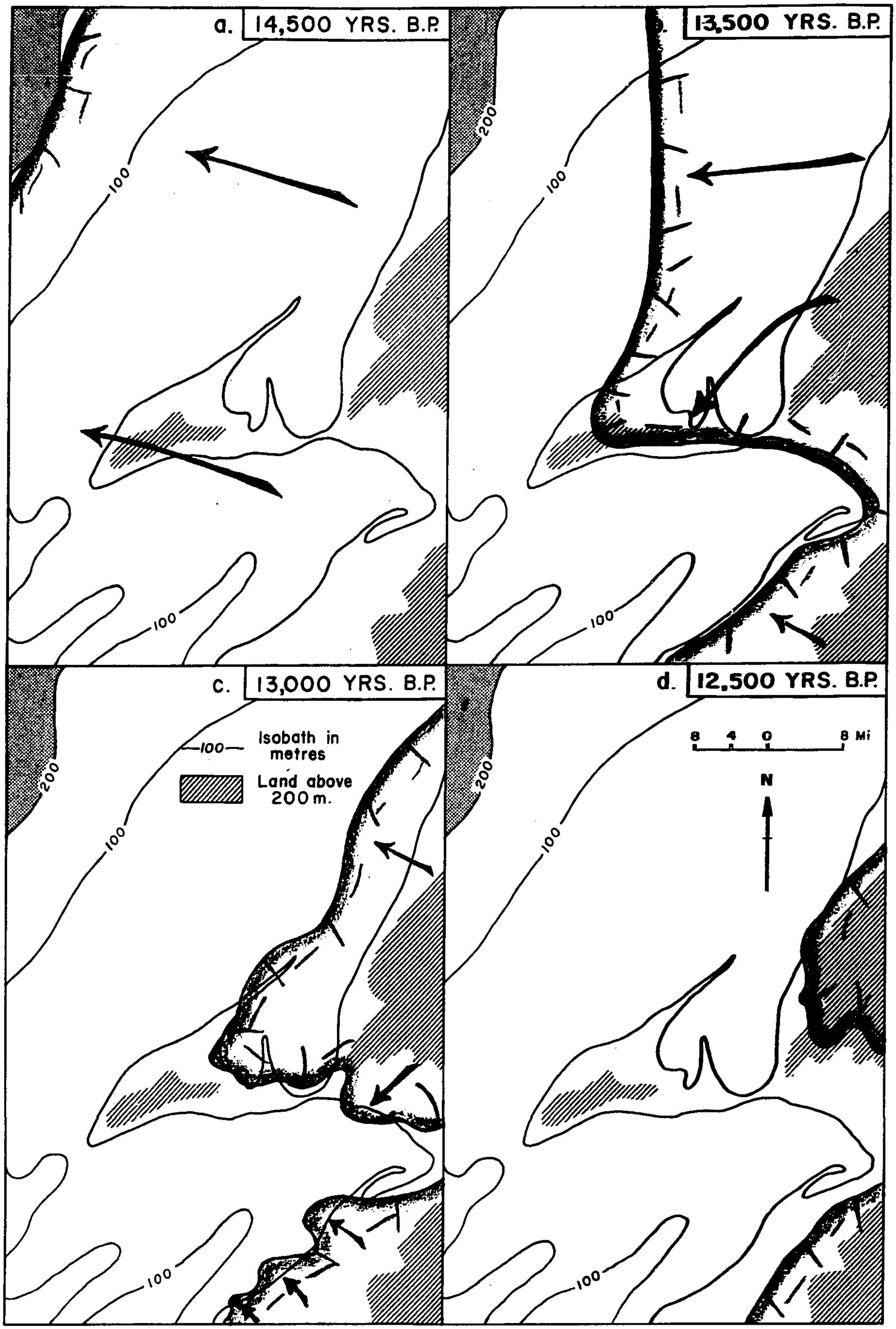


Figure 61: Speculative ice-marginal positions during the deglaciation of southwestern Newfoundland and the off-shore zone

before that time. Ice flow would have been at least as active in the latter area as in the fjords. Marine landforms in the area south of Serpentine River have been found up to an elevation of 130 feet above sea level. A 30-foot high bluff, the base of which is at 130 feet, is cut into the seaward edge of a fan of debris deposited in front of an end moraine at the mouth of Rope Cove canyon (figure 62). This end moraine is contiguous with that built in contact with an ice lobe in the trough now occupied by Serpentine Lake.

In order to provide a tentative date for these morainic features, reference must be made to the reconstruction of the regional trend of isostatic warping (figure 60). That figure shows that there is a difference of about 40 feet in the hypothetical elevations of marine limit features at Cox's Cove, Bay of Islands and the area south of Serpentine River. These elevations are those which the features would have today if they had been formed at 13,700 years B.P. There is a difference in the actual elevations of marine limit features in these two areas of about 30 feet, which suggests that they are comparable in age. Since a radiocarbon date of 12,600 ± 170 years B.P. (G.S.C.-868) is available for the features in the Bay of Islands, a similar age can be assigned to those seen south of Serpentine River. Those end moraines are of the same age and, therefore, are comparable to the fjord-head stand of glaciers in the Bay of Islands.

ii. Other terraces. In the fjord-head and fjord-side localities in the Bay of Islands and Trout River where marine overlap was examined, another prominent regressional terrace lies below the marine limit. At Cox's Cove, this stands at 85 feet above sea level, or 75 feet below the marine

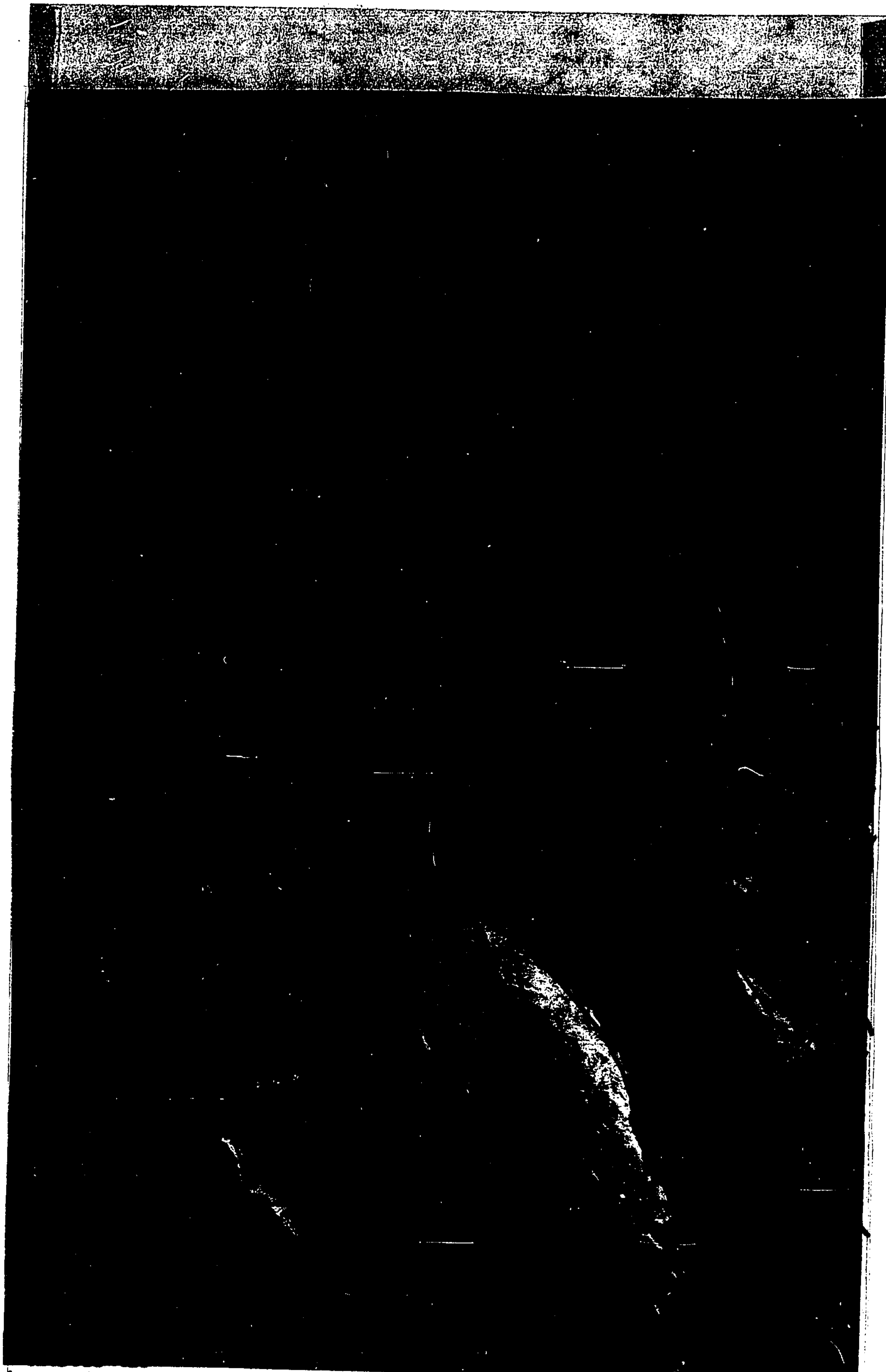
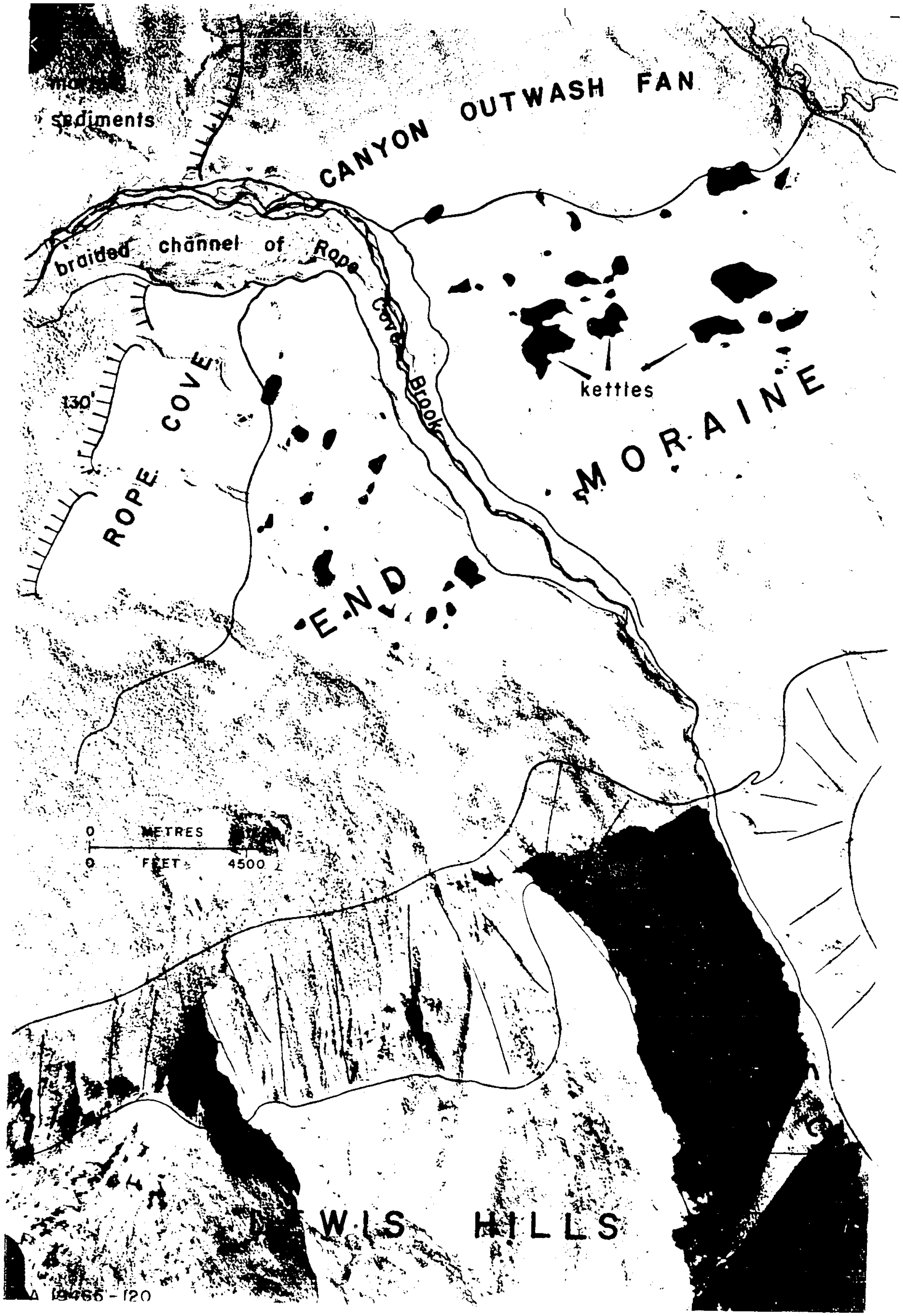


Figure 62: Landforms at the mouth of Rope Cove Canyon.



sediments
CANYON OUTWASH FAN
braided channel of Rope Cove Brook
kettles
ROPE COVE
130'
END MORAINE
0 METRES
0 FEET 4500
WIS HILLS
A 19466-120

limit. At Trout River, a delta terrace at 115 feet stands 80 feet below the highest discernible marine feature. Since Cox's Cove was not deglaciated until 12,600 ± 170 years B.P. (G.S.C.-868), and a comparable date may be given to that event at Trout River, these terraces cannot be the same age as those seen at 85 to 90 feet around St. George's Bay. According to the model of land- and sea-level changes at Cox's Cove, sea level would have fallen to 85 feet at 11,800 years B.P., and at Trout River it would have fallen to 115 feet at the same date. The date of 11,800 years B.P. for the 85-foot and 115-foot terraces at Cox's Cove and Trout River, respectively, is the same as that at which sea level passed below present datum at Abrahams Cove (figure 58).

Relatively little attention was paid to lower regressional features in this study. Flint (1940) collected a great deal of data on such features between Port au Port Bay and Bonne Bay, and drew isobases on a surface which included them. He called this the "Bay of Islands Surface," and showed it sloping up in the direction N20°W, from a zero isobase through Stephenville, to 100 feet through the head of Bonne Bay (Plate V). The direction of tilt on this surface is approximately the same as that on the surface Flint reconstructed on the highest observed marine features. I have proposed that tilt actually slopes up towards the northeast, rather than the north-northwest. It is difficult to see how a lower, therefore later, and presumably synchronous surface, such as the "Bay of Islands Surface," could have such a different direction of tilt.

There is cause to call many of the elevations shown in Flint (1940) into question. In cases where I am able to agree with his identification of a feature, nowhere do I agree with the elevation he has shown for

Figure 63: Trout River village, viewed towards the west from the road to Bonne Bay. Upper marine limit at 195 feet in foreground; main "115-foot" delta terrace in middle distance, and lower, rock-cut platforms above headland (see figure 47).

Figure 63: Trout River village, viewed towards the west from the road to Bonne Bay. Upper marine limit at 195 feet in foreground; main "115-foot" delta terrace in middle distance, and lower, rock-cut platforms above headland (see figure 47).

it. Flint's elevations are usually underestimated by up to ten feet. Often I have also recorded features higher than the highest which he shows at a locality where several features are present. Since all elevations in this study upon which significant conclusions are based have been fixed with an engineer's level from a datum fixed using tide tables, I see little possibility of accepting the majority of Flint's measurements. However, since Flint (1940) shows about eighty localities (some with more than one feature recorded) for which height determinations on marine features closely approximating sea levels were made, it can at least be expected that his errors, whatever their magnitude, are systematic.

I feel the difference of opinion can be resolved if the following points are considered. Firstly, there is little evidence in Flint's data upon which a zero isobase on his "Bay of Islands Surface" can be fixed. Secondly, the 50-foot isobase on this surface passes through features on the sides of Humber Arm, Bay of Islands, but no features support this line on the open Gulf shoreline. Further, in the Bay of Islands, Flint shows only one feature at any one site, whereas more than one is usually present, both above and below the estimated elevation he shows. Recognition of these other features would upset the regularity of isobases on his "Bay of Islands Surface." Further, in the Bonne Bay area, Flint recognizes the profusion of features to be seen there (although he omits some prominent ones), but does not justify the choice of features near 100 feet which are used to draw the 100-foot isobase in that area. At Trout River, for instance, where Flint shows features at 158, 98-103, 100-108 (these two probably being the same feature in two different places), and 60 feet, I have levelled features on three sides of this semi-circular cove, and they can be grouped into levels at 195, 180, 157, 130, 115, 105, 85, 75-79, 69, 47, and 35 feet (Plate VIII and figure 63). The 115-foot level is the most prominent regressional feature, yet Flint's 115-foot isobase lies eleven miles north of Trout River. Finally, some of Flint's data can be accommodated on a surface sloping up to the northeast, similar to that reconstructed in figure 60.

The foregoing reconsideration of Flint's reconstruction of a tilted plane embracing regressional marine features has pointed to the difficulties encountered in accounting for the variation in the elevations of these features. Although less attention has been paid to them in this study than to the marine limit features, I feel that two tentative conclusions can be drawn. Firstly, some of the altitudinal variation is due to a differential isostatic tilt which has occurred since about 11,800 years ago, when the most prominent regressional features north of St. George's Bay were formed. This tilt cannot depart significantly in direction from that tentatively assigned to the hypothetical marine limit plane formed at 13,700 years B.P. Secondly, other variations in the elevations of regressional marine features are partly attributable to their erratic distribution in isolated coves, where conditions favourable for their formation prevailed. The possibility of the occasional pre-Wisconsin terrace disturbing the pattern must also be considered. Particularly, at Tea Cove on Port au Port Bay, resistant sandstones are planed horizontally for a quarter of a mile inland from coastal cliffs, and this surface, at about 45 feet, is overlain by angular gravels in a sandy matrix. While the gravels may well be postglacial, it is difficult to assign a postglacial age to this rock-cut terrace with the rapid rate of postglacial sea level lowering shown in the model for Abrahams Cove (figure 58).

b. Radiocarbon-dated horizons below the marine limit

Two radiocarbon dates have been obtained on organic material below the marine limit in southwestern Newfoundland. Discussion of these has been postponed until this stage because no definite conclusions could be reached earlier as to the relation of the dates determined to the elevation from which the materials were recovered.

A date of 7,340 ± 220 years B.P. (G.S.C.-1145) was obtained for plant detritus removed from compacted sand at the base of a peat bog at Turf Point, St. George's (figure 64). The location of the site, and the stratigraphy in the area are shown in figure 65. The village of St. George's lies on an eroded slope of St. George's River Drift till. This till overlies red Mississippian sandstones which outcrop in low ledges along the shore south of Turf Point. The till slope is marked by faint regressional shorelines and loosely consolidated sands overlie the flats which represent minor stillstands of sea level. Faint terraces at 26, 35, and 85 feet have been levelled in this study. The 85-foot level is the wave-worn counterpart of a terraced delta surface which is prominent to the north and south of St. George's. The deposition of the original deltaic wedge was interrupted along the St. George's shoreline because the post-glacial streams, of which Little Barachois Brook and Flat Bay Brook are descendants, did not deposit sediments in that area.

Bog growth along the St. George's shoreline could not have begun until sea level had been lowered below its present stand. The curve of relative sea level at Abrahams Cove shows that this took place at 11,800

Figure 64: Wave-eroded exposure in peat bog at St. George's. C^{14} dating sample from basal sand. Note wave-washed boulders from underlying till.

Figure 66: Submerged baymouth bars inland of modern bar, Stephenville Pond, Harmon Field.

Figure 64: Wave-eroded exposure in peat bog at St. George's. C^{14} dating sample from basal sand. Note wave-washed boulders from underlying till.

Figure 66: Submerged baymouth bars inland of modern bar, Stephenville Pond, Harmon Field.

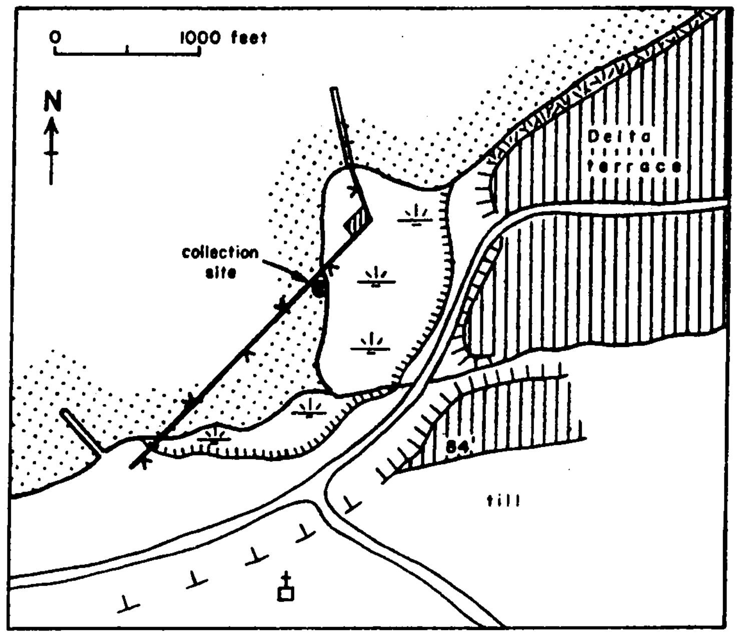
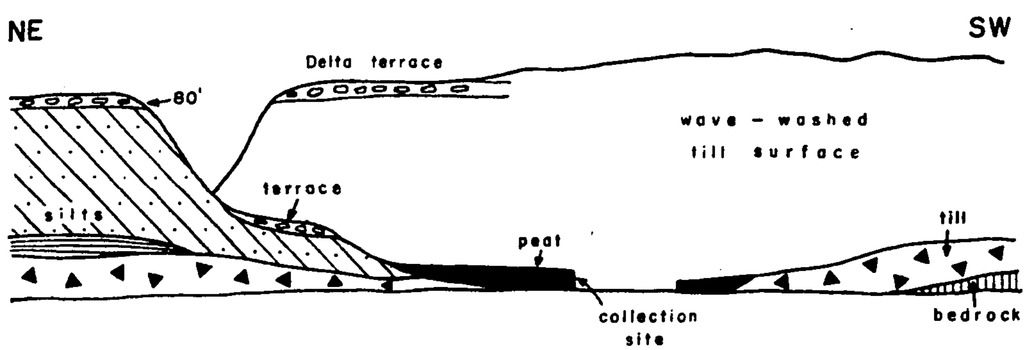


Figure 65. Location and stratigraphic position of peat bog, St. Georges.



years ago, but also that sea level continued to fall until about 10,000 years ago. An 800-year period of maximum land emergence was terminated at 9,200 years B.P., after which sea level began to rise towards its present stand. The date of 7,340 years B.P. on the basal material of the St. George's bog can merely be interpreted as the date at which bog deposition began. The site would have been well-drained following the fall of sea level below the modern datum; in fact it would have been increasingly well-drained until the period of maximum emergence. Bog deposition would begin when drainage became impeded, and this would have occurred when sea level had risen far enough for the water table at St. George's to be near present sea level. At 7,340 years B.P., the Abrahams Cove model shows that sea level was just above minus 5 metres (about 16 feet below present). That bog development can progress on level areas this close to sea level around the present shoreline in southwestern Newfoundland is shown by the wide expanse of bog that covers Shoal Point, separating East Bay and West Bay, Port au Port Bay. There, bog growth of the kind seen at sea level at St. George's is active at 20-30 feet above sea level. Therefore, while the date on the St. George's bog could not confidently be used in the derivation of a model of sea-level change, the latter provides the basis for confidence in the date and the interpretation made of environmental conditions. That the bog is presently being eroded by waves indicates that conditions favourable to its development once extended further seaward. Since the sea floor slopes gently offshore, it is likely that the bog thickened seawards and that the exposed section is not the maximum thickness. The date obtained at the base may, therefore, not be the earliest which could have been obtained from an undisturbed bog in this position.

A date of 5,760 ± 210 years B.P. (G.S.C.-1203) was obtained on shell fragments in sands in a submarine core taken from West Bay, Port au Port Bay.[2] The core was taken in 11 fathoms (66 feet) of water, and the sandy zone extends from 210 to 220 cms. below the top of the cores. The shells, therefore, are at an elevation of minus 70 feet (21 metres). Shearer (personal communication, 1969) interprets the sandy zone as representing a period of maximum emergence and, therefore, minimum depth of water in West Bay. His calculations of the minimum depth of water in which orbital velocities would be capable of moving sand grains of the size found in the shelly zone led him to the conclusion that a maximum emergence of 35 to 45 feet had occurred in the bay. This estimate is close to that obtained from the curve of sea level change at Abrahams Cove (figure 58). However, the date of 5,760 years B.P. is 3,500 years younger than the date at which sea level last stood near - 35 feet, according to the Abrahams Cove sea level curve. Possibly, the shells which Shearer recovered have been emplaced at a level below that at which the sea stood at the time of their death. This might be accounted for as the result of sea-floor scouring and filling which would be more powerful after maximum emergence. Flow-tide currents of eight knots are presently experienced in West Bay.

c. A Note on Contemporary Sea-Level Movements

With regard to the most recent history of changing land and sea level, little evidence was obtained in this study. The models of land-and

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2. The shell sample was collected by J. M. Shearer, Dalhousie University, Halifax, in the course of his investigation of the bottom sediments of Port au Port Bay. I am grateful to him for allowing me to include the material in the discussion.

sea-level change, derived in section ii of this chapter are insufficiently accurate for the last 5,000 years. This is due to the fact that the slow rates of uplift and eustatic sea-level change in that interval provide only wide time limits for the determination of the ages of features such as bedrock shore platforms and coastal bars and spits. Also, the detailed form of eustatic sea-level changes is insufficiently known for the last 5,000 years. Further, the effects of hydroisostatic subsidence have not been gauged.

More positively, it can be noted as a concluding remark that the presence of coastal peat bogs undergoing wave erosion suggests that sea level is currently rising in southwestern Newfoundland. Grant (personal communication, 1970) has discovered a small remnant of a drowned forest on the south shore of East Bay, Port au Port, at Boswarlos. He tentatively places the dividing line between submerging and emerging littorals in western Newfoundland near the base of the Great Northern Peninsula of the island. At Port Harmon, near Stephenville, a series of baymouth bars can be seen declining in elevation inland of the modern beach, beneath the waters of Stephenville Pond (figure 66). This is a further indication of recent submergence. The extension of work on coastal submergence in the Maritime Provinces (Grant, 1970a) to western Newfoundland is an important and enticing topic for future research.

CHAPTER EIGHT

A Resume of Glacial Events in Southwestern Newfoundland

This chapter is designed to present a unified picture of the sequence of glacial events which occurred in the later part of the Wisconsin age in southwestern Newfoundland. Some reference is made to recent studies in the Maritime Provinces, the results of which are of signficance to those obtained here. Also, several outstanding problems are raised which are intended as pointers to future research.

The earliest event for which evidence is available is that in which an ice sheet covered the entire area, to an undetermined height above the highest summits at 2,500 feet, and to an undetermined distance beyond the present shoreline. An hypothetical ice surface profile suggests that those summits were covered by less than, and probably much less than, 1,000 feet of ice. The edge of the Laurentian Channel, 35 to 70 kilometres (22-44 miles) offshore and at about 100 fathoms (ca. 200 metres) below sea level, can be suggested to have been the western limit of this ice sheet.

The westerly extent of Newfoundland ice at the glacial maximum could be ascertained by bottom sounding techniques which allow tills to be distinguished from other sediments. Further, the distinctive petrography of some western Newfoundland rocks could be traced in the tills which doubtless underlie the bottom sediments for some distance offshore. Recently, cores taken from the floor of the Laurentian Channel have revealed thin, brick-red "till" layers between layers of glacio-marine sediments (Connolly, Needham, and Heezen, 1967). The colour and petrography of the "tills" was interpreted by these workers as evidence of a derivation from red beds of Permo-carboniferous and Triassic age in the Appalachian-Acadian region. Doubt may be expressed as to whether the brick-red layers are in fact tills or till-like materials dropped from the undersides of floating ice masses during the late-glacial marine invasion of the Gulf of St. Lawrence (Grant, personal communication, 1970). It has been suggested in this study that the great depth of the Laurentian Channel permitted a more rapid marine invasion along its arms than over gently shelving zones inshore of these. The tentative age of 15,000-11,000 years B.P. which the aforementioned authors assigned to the "till" layers prompts the suggestion that the materials might be ice-rafted.

Glacial striae and erratic till boulder provenances in the area of Port au Port and St. George's Bays indicate that ice flowed westwards and north of westwards at the last glacial maximum from outflow centres over the Long Range Mountains. During the easterly retreat of the margin of the ice sheet in the Gulf of St. Lawrence, ice flow directions became more variable. This was due to the relatively greater influence of ice flow from the Bay of Islands Mountains, standing at 2,500 feet to the west of the 2,000-foot summits of the Long Range Mountains, and to the changing orientation of the ice margin in the offshore zone that was induced by the calving action of late-glacial sea waters flooding the Gulf, first up the deep arms of the Laurentian Channel, later over shallower inshore zones, and finally along the fjords.

During the glacial maximum and its retreat phase, a sheet of massive, compact, lodgement till was deposited beneath the ice sheet. Isolated pockets of ice-contact stratified debris above the till indicate local stagnation along the present shoreline immediately prior to marine overlap. Both these facies are grouped together as the St. George's River Drift.

There is no evidence in southwestern Newfoundland, from ice flow features, till provenances, or the regional trend of isostatic warping, that Labrador ice had any effect on the glaciation of the area. Prest and Grant (1969) have argued the case for the dominance of locally nourished ice-cap complexes in the glaciation of the western Gulf of St. Lawrence and the Maritime Provinces. They viewed the Laurentian Channel as drawing down the undoubtedly strong flow from Labrador southeastwards towards Cabot Strait. The same can be argued for western Newfoundland and its offshore area. While Labrador ice is not in evidence in southwestern Newfoundland, Grant (1970b) has found evidence of ice flowing vigorously from Labrador across the northern part of the Great Northern Peninsula of the island, prior to a westward ice advance off the Long Range Mountains. The sea floor between Newfoundland and Labrador is relatively shallow and smooth and consequently offered no barrier to southeastward-moving Labrador ice. South of Port au Choix, however, Newfoundland ice seems to have dominated the flow into the Gulf because Labrador ice was drawn southwestwards along the northeastern arm of the Laurentian Channel, known as Esquiman Channel. (Plate I).

The present shoreline of southwestern Newfoundland was first overlapped by late-glacial marine waters at about 13,700 years B.P. The earliest date for overlap comes from the south shore of the salient Port au Port peninsula, which lies closest to the Laurentian Channel. However, dates of 13,400 years B.P. from Port au Port village, and 13,500 years B.P. from the southern extremity of St. George's Bay show that overlap was approximately synchronous all around that bay. Port au Port Bay appears to have been deglaciated somewhat later, between 13,000 and 12,600 years B.P., due to the isolation in the bay of a lobe of lowland ice. The fjord valleys of the Bay of Islands were not ice-free until 12,600 years B.P., by which time marine waters had flooded their entire length. The topographically similar valleys of Bonne Bay and Trout River Ponds were, by analogy, also deglaciated at that time.

The radiocarbon dates presently available for the initiation of late-glacial marine overlap in southeastern Canada show that the event occurred within a relatively short time interval (figure 57). The date of 13,700 ± 230 years B.P. (G.S.C.-1074) from Abrahams Cove, western Newfoundland, is the earliest recorded for the event northeast of the Connecticut shoreline, and that of 12,600 ± 170 years B.P. (G.S.C.-868) from Cox's Cove, western Newfoundland, is the second youngest from the region. Presently, the northeast coast of New Brunswick, the north coast of Newfoundland, and the north shore of the Gulf of St. Lawrence are areas from which no radiocarbon dates are available on the start of marine overlap. Dates need to be collected from these areas in order to provide a basis for the assessment of the influence of rates of isostatic uplift and different dates of ice retreat from the present shoreline upon the progress of marine overlap over the entire Atlantic region of Canada.

At Abrahams Cove, western Newfoundland, the similar dates for shells recovered from the base and from the top of the marine deposits, separated vertically by 120 feet, indicate that marine overlap occurred instantaneously to depths of over 100 feet. This was possible because the area was isostatically depressed beneath a eustatically rising sea level which stood high on the outer edge of the glacier prior to overlap. Rapid rebound was causing sea level to fall relative to present, and marine limit features were registered against the present shoreline as soon as ice cleared a site. Till clasts in shelly basal marine clays attest to the presence of drifting bergs seaward of the ice margin during the early stage of overlap, while an upward coarsening of deltaic sands and gravels indicates a rapid influx of meltwater into the sea from streams draining an ice front which stood only a short distance inland. The deltaic sequence is known as the Bay St. George Delta.

In several coastal cliff sections around St. George's Bay deposits of a later glacial event are seen in intimate association with the upper part of the Bay St. George Delta sequence. This relationship is the result of local resurgences of the ice margin into the sea following its wastage a short distance inland of the present shoreline. Between the lobate surges, the ice margin merely halted temporarily in its retreat into the mountains surrounding the bay. This interpretation of the glacial event which produced the Robinsons Head Drift denies it any regional significance in the present landscape. The event is tentatively dated at 13,000 years B.P. with the aid of the model of sea level change. At this time, fjord valleys further north were still ice-bound, and any effect the event may have had in offshore areas outside St. George's Bay has been hidden or obliterated by subsequent sea-level rise.

A slightly later stand of the ice margin is marked by the end moraine at the western end of Serpentine Lake and by the fjord-head stand of valleys glaciers in the Bay of Islands and Bonne Bay. Following this stand, the only other definable glacial event in the general retreat into the highlands is marked by prominent cirques that are particularly well developed around the margins of the Bay of Islands Mountains.

Upper marine limit features in southwestern Newfoundland stand at elevations between 113 and 150 feet around St. George's Bay and on the Gulf shore of the Port au Port peninsula. In Port au Port Bay late ice prevented the registration of marine features until sea level had been drawn down to about 50-55 feet. In the Bay of Islands the limit stands at 160 feet, and in the Bonne Bay-Trout River area at 195 feet. While the general rise in these elevations along the southwest to northeast trend of the coastline indicates a component of isostatic tilt in that direction, the radiocarbon dates on the marine limit in different topographic situations indicate that its variable age has added a complicating effect to this pattern.

No dateable regressional marine features were recorded which would have permitted an uplift curve to be derived for any site. However, the theoretical considerations of Andrews (1968a) have permitted tentative uplift curves to be drawn for a site on an open coast and for another in a fjord situation, for both of which an elevation and a date for the marine limit are available. These curves, considered with a curve of eustatic sea-level rise, enable a trace to be drawn of the position of sea level through time relative to its present datum. On this trace, sea level at Abrahams Cove was falling at a rate of about nine feet per century in the 1,000 years following deglaciation. From its marine limit stand, sea level at this site had reached present datum at 11,800 years B.P., and continued to fall to about 35 feet below present, a position it occupied 10,000 to 9,200 years ago. Since the latter time sea level in St. George's

Bay has been rising to its present stand.

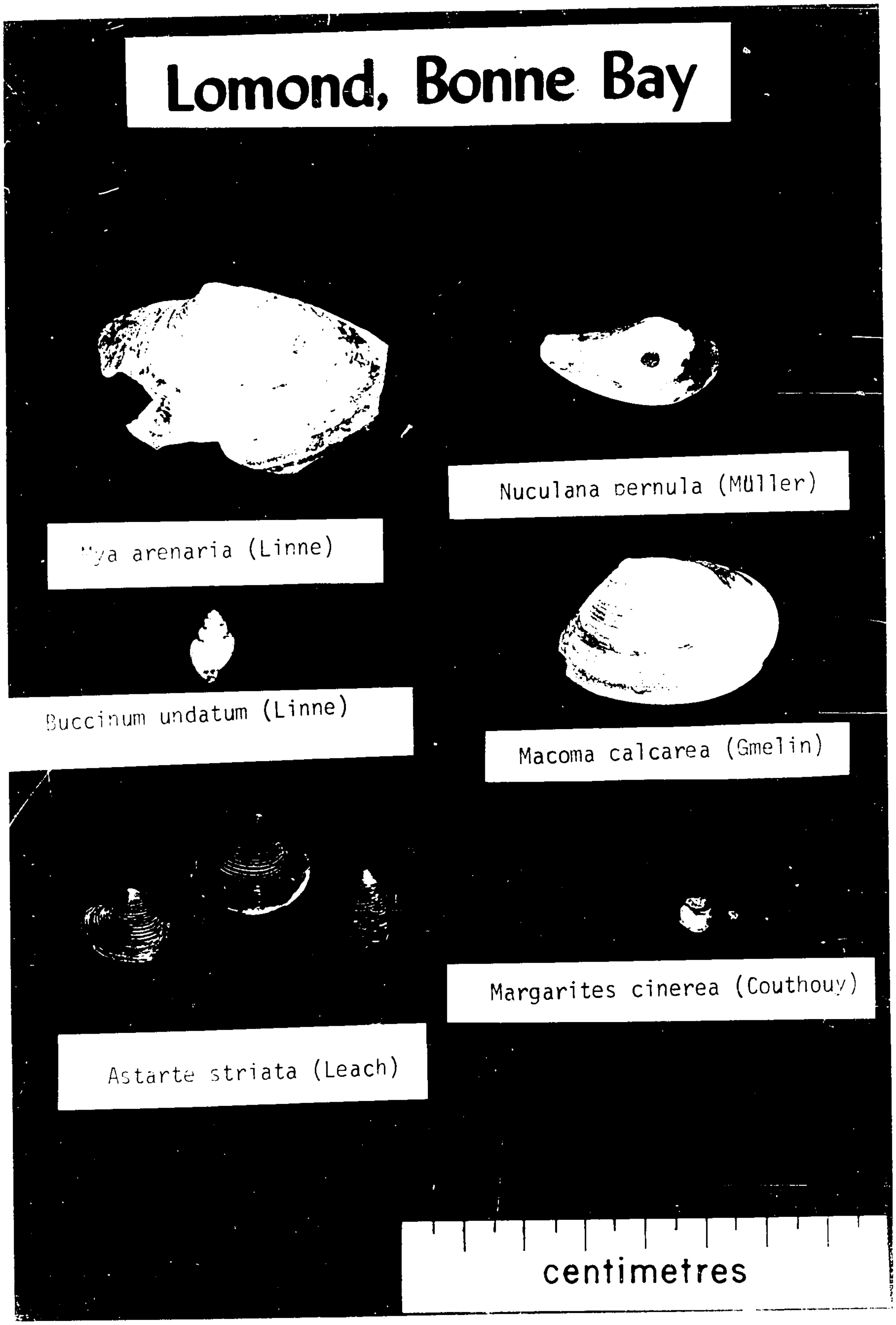
The uplift curves also make it possible to estimate the extent to which the variation in marine limit elevations is due to their different ages and to differential warping. The hypothetical elevations of features were calculated on the assumption that they had formed at the same time as the earliest of them - the Abrahams Cove feature, 13,700 years old. These elevations, on an hypothetically synchronous sea-level plane, permit the direction of isostatic warping to be determined. The triangle method shows that this warping is up towards the northeast at two feet per mile: nearly at right angles to that shown by Flint (1940). This trend most likely results from the strong influence of a Newfoundland ice cap on crustal depression at the glacial maximum.

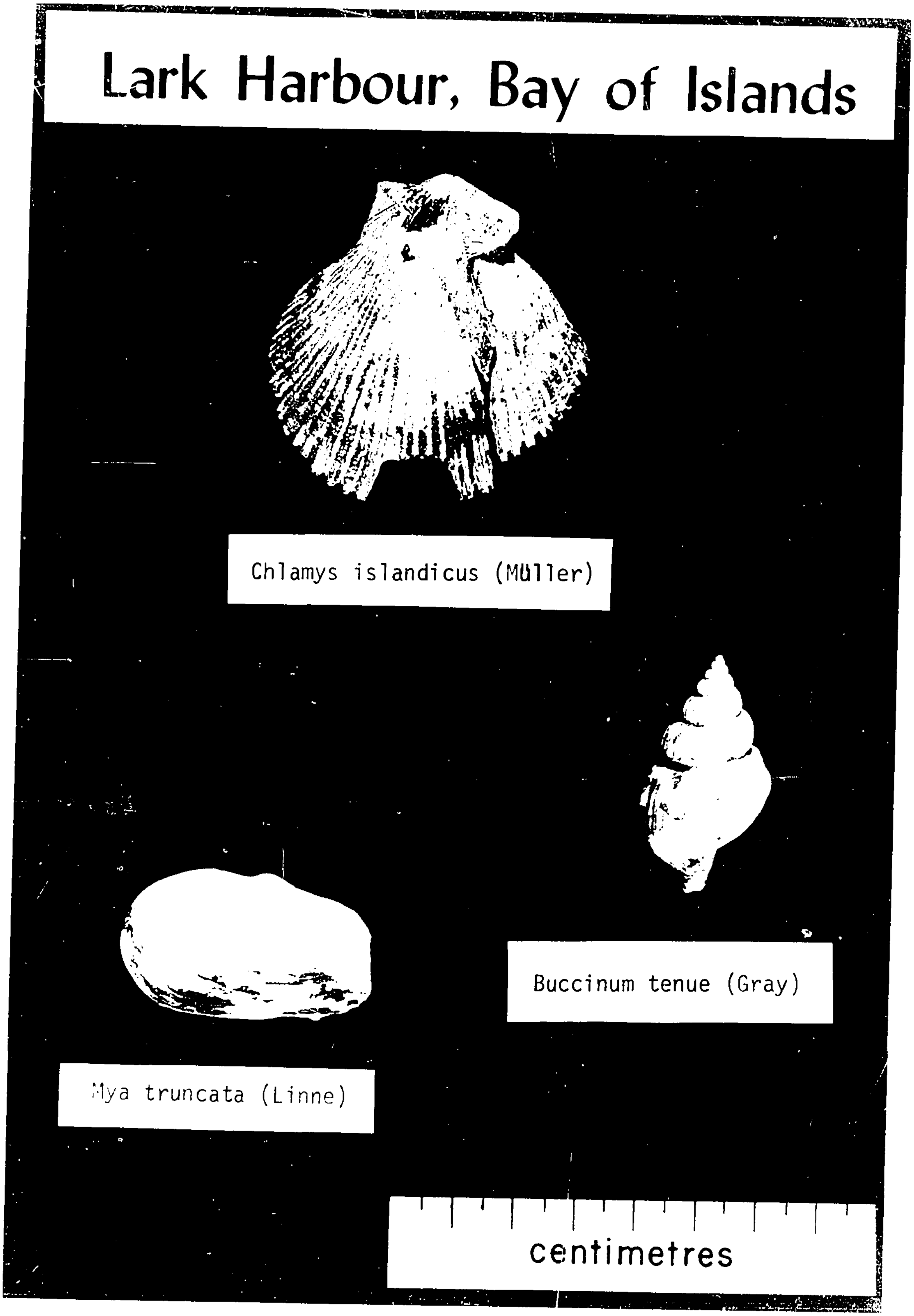
The curves of relative sea-level change are also used to assign tentative ages to regressional features, such as the terrace formed penecontemporaneously with the Robinsons Head Drift glacial event at 13,000 years B.P. The major regressional features in the Bay of Islands and Bonne Bay-Trout River are tentatively assigned an age of 11,800 years B.P., a time at which sea level stood at its present level in St. George's Bay.

Further research is essential in Newfoundland and all other areas around the Gulf of St. Lawrence in order to ascertain directly the regional trends of isostatic warping. Isobases can only be drawn on synchronous features, and it might be expected that there would be few areas in which they were formed at the same time because of the local topographic variation on land and offshore areas. If uplift curves with close dating control become available, sites may be compared with respect to the amounts of uplift which had occurred at specific dates, and graphical methods can then be employed to determine warping trends. Particularly, in Newfoundland these studies would assist in ascertaining the degree to which the dimensions of the island ice cap affected glacio-isostatic warping.

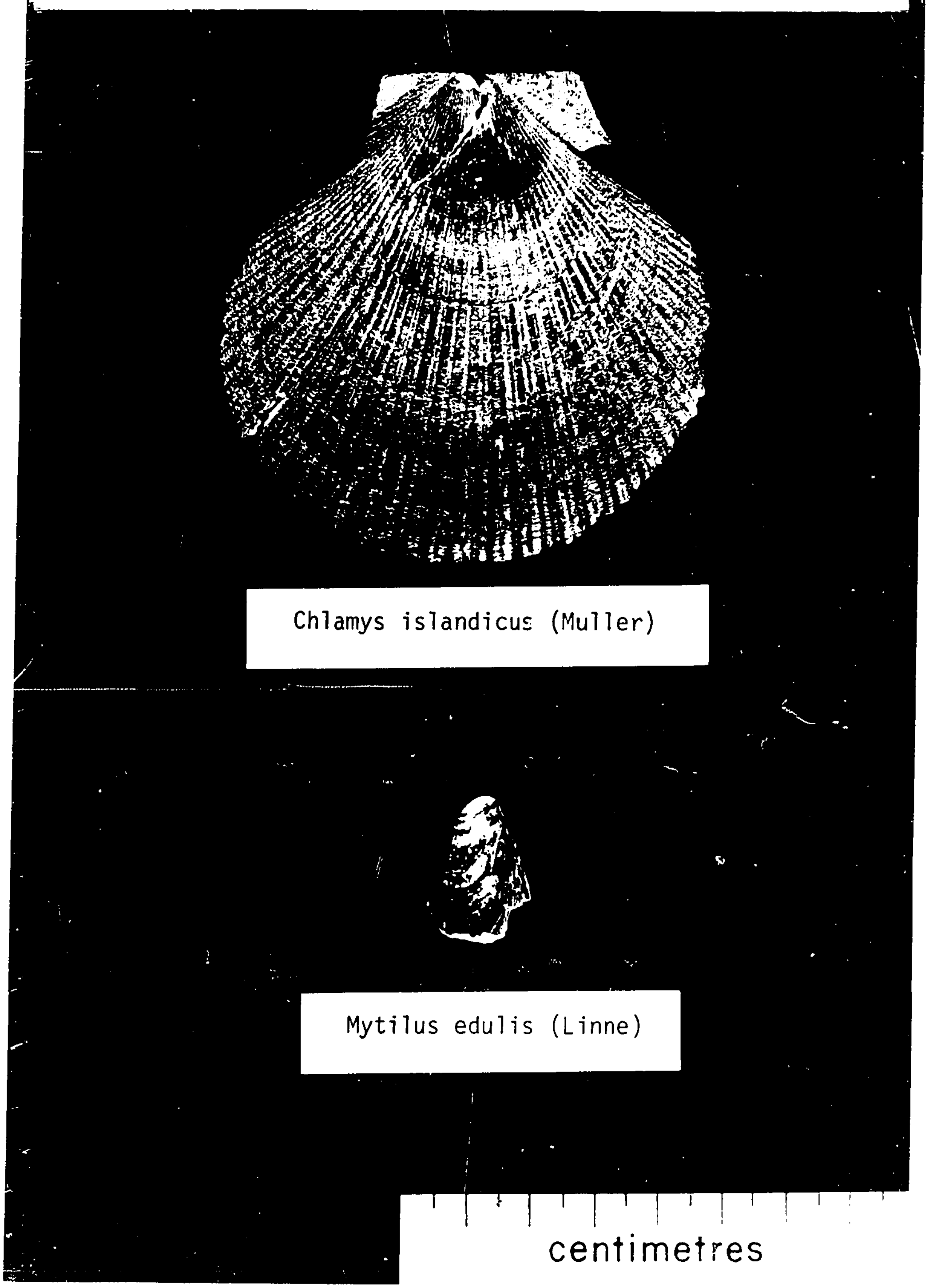
APPENDIX A

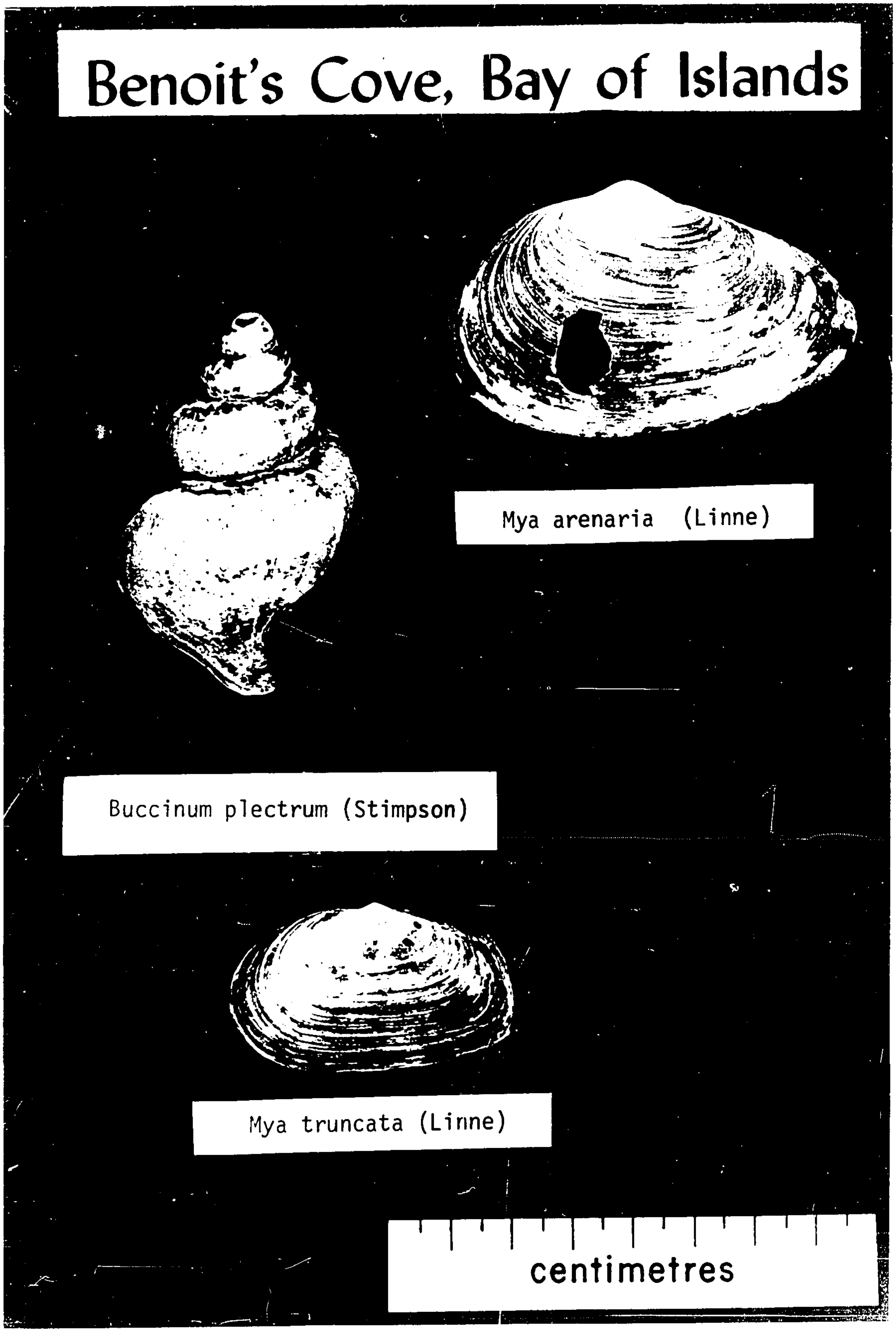
An Illustrated Catalogue of Pleistocene Marine Macro-Fauna from Selected Sites in Southwestern Newfoundland


Lomond, Bonne Bay
Nuculana pernula (Müller)
ya arenaria (Linne)
Buccinum undatum (Linne)
Macoma calcarea (Gmelin)
Margarites cinerea (Couthouy)
Astarte striata (Leach)
centimetres

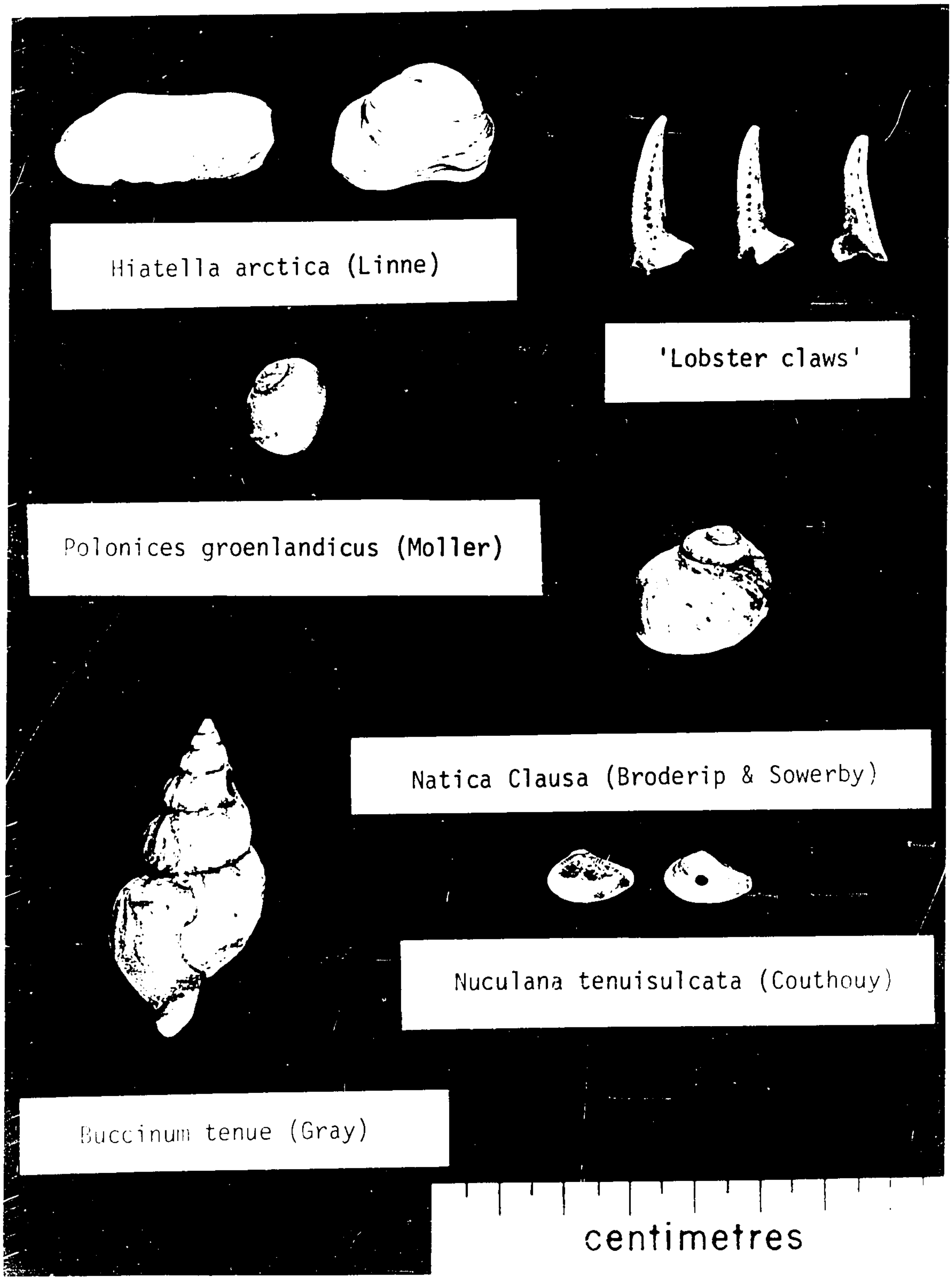

Lark Harbour, Bay of Islands
Chlamys islandicus (Müller)
Buccinum tenue (Gray)
Mya truncata (Linne)
centimetres

Benoit's Cove, Bay of Islands

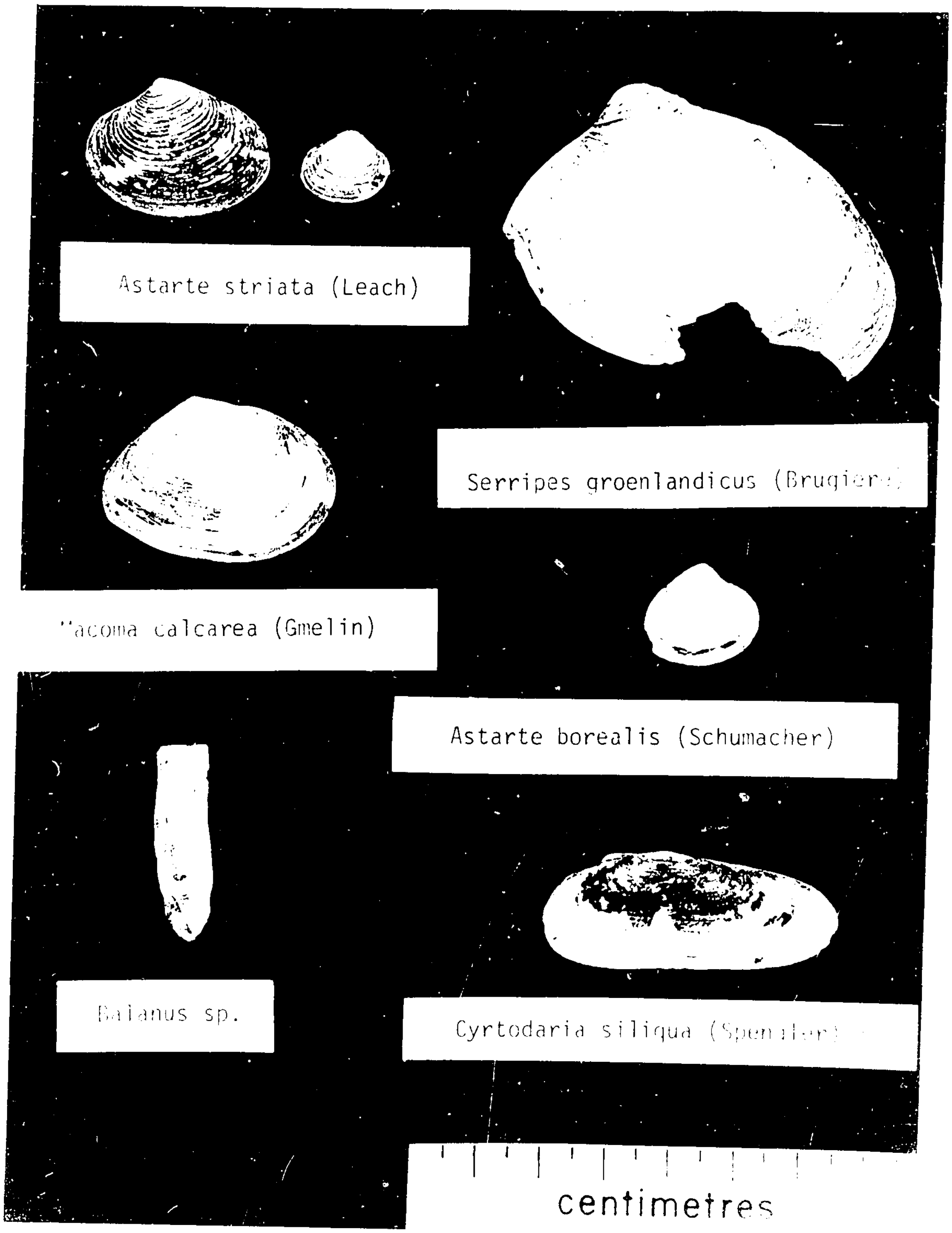


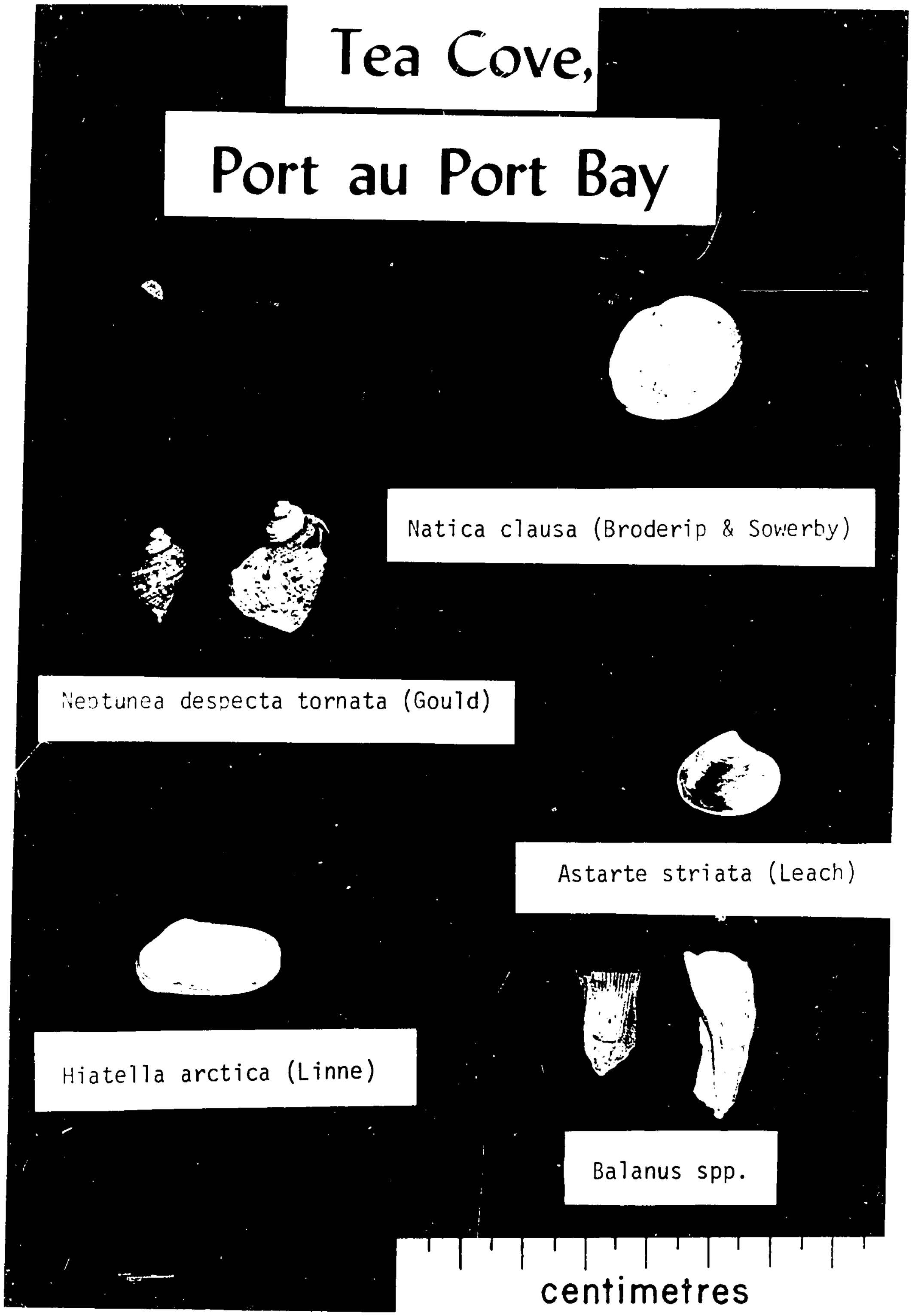

Benoit's Cove, Bay of Islands
Mya arenaria (Linne)
Buccinum plectrum (Stimpson)
Mya truncata (Linne)
centimetres

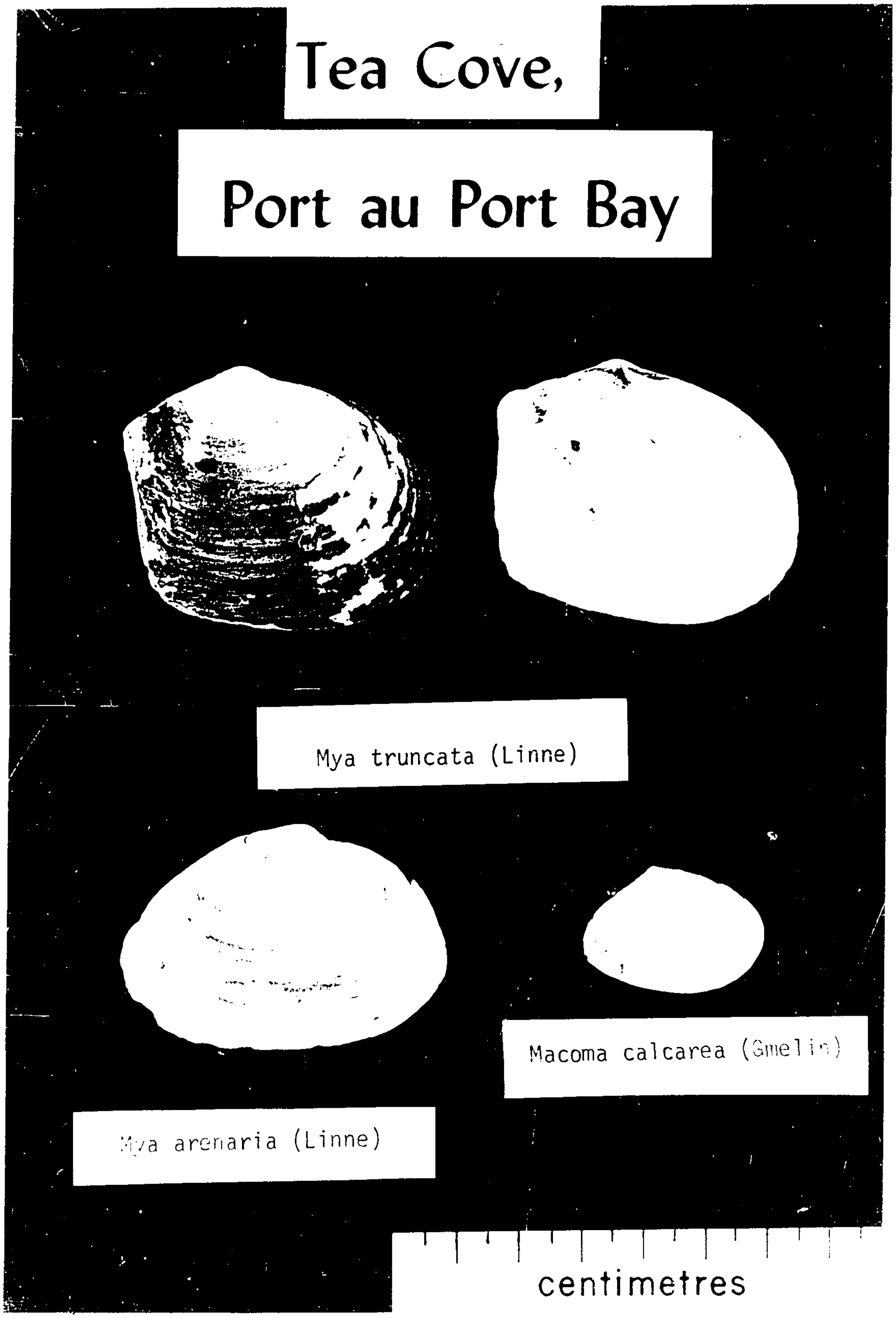
Benoit's Cove, Bay of Islands



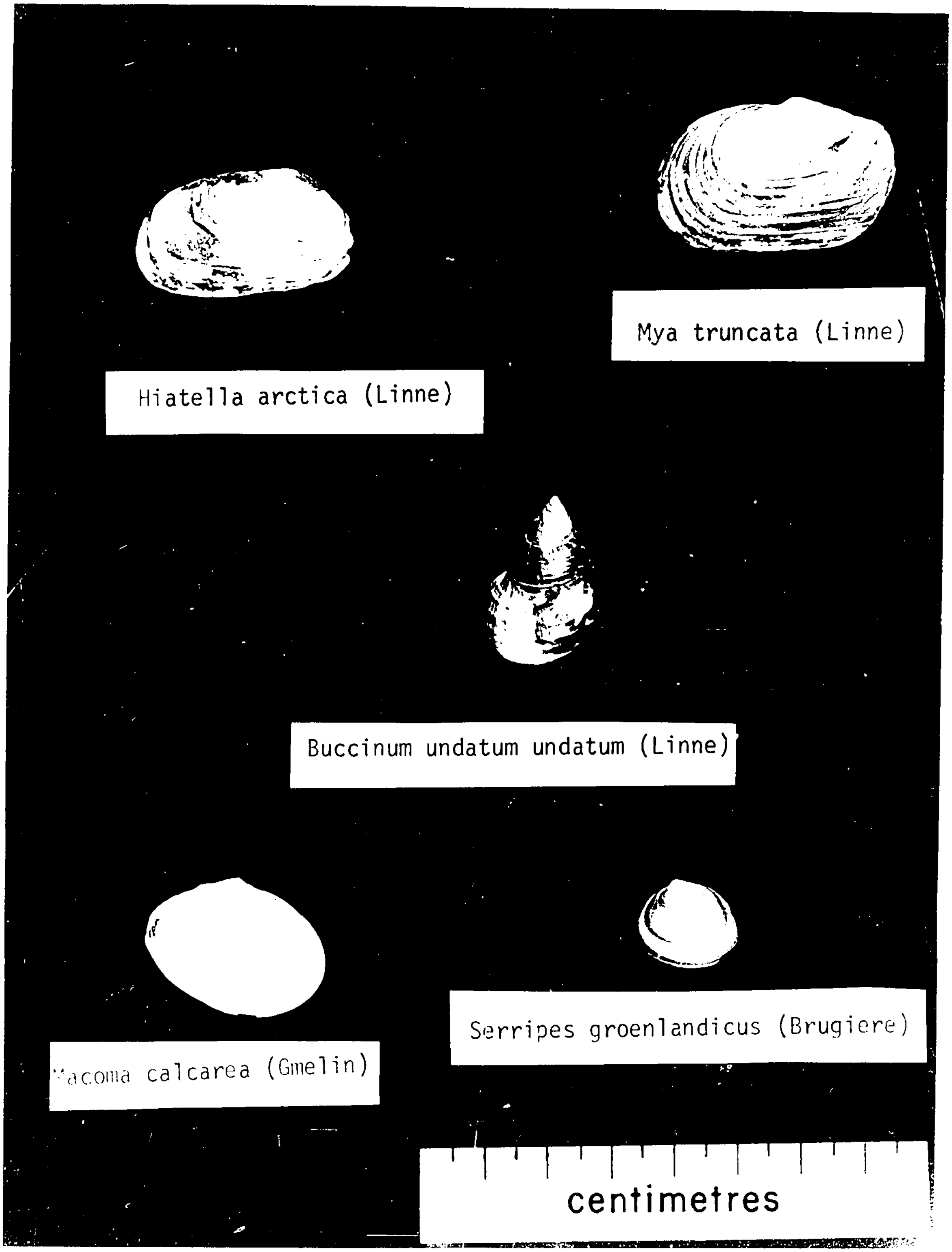
Benoit's Cove, Bay of Islands




Tea Cove,
Port au Port Bay
Natica clausa (Broderip & Sowerby)
Neptunea despecta tornata (Gould)
Astarte striata (Leach)
Hiatella arctica (Linne)
Balanus spp.
centimetres


Tea Cove,
Port au Port Bay
Mya truncata (Linne)
Mya arenaria (Linne)
Macoma calcarea (Gmelin)
centimetres

Abrahams Cove, St. George's Bay



Butter Brook, St. George's Bay

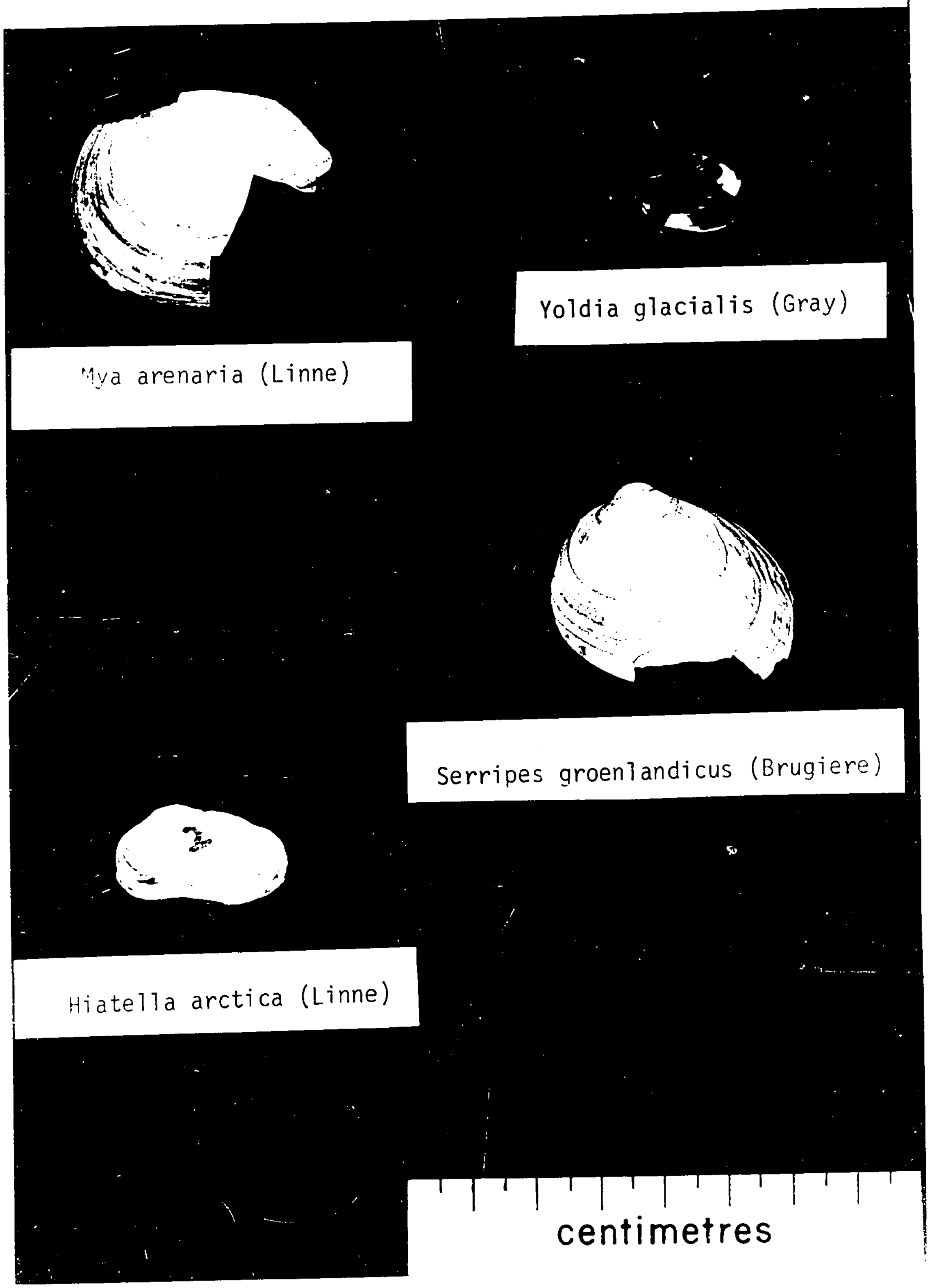


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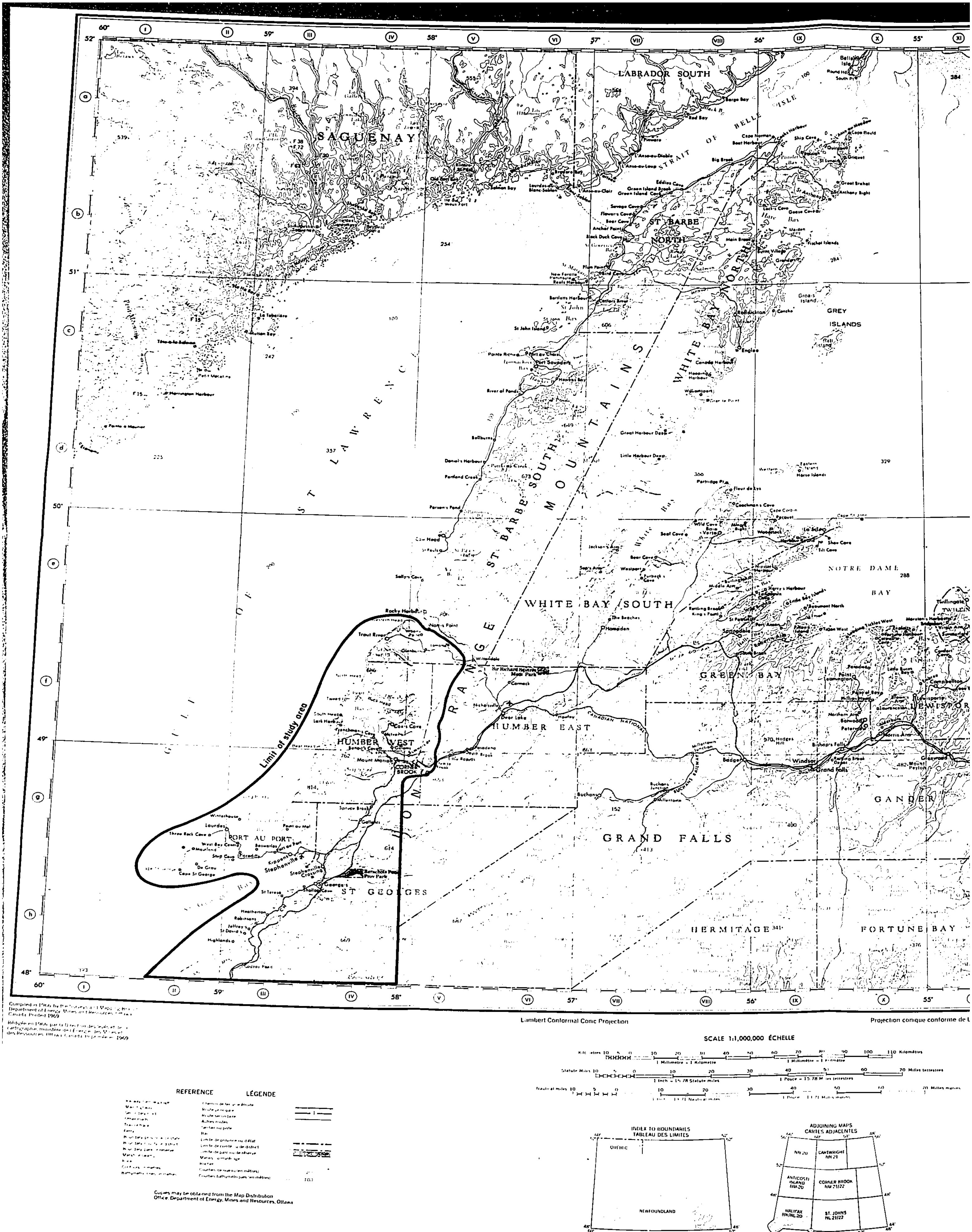
PLATE I

[redacted]



LABRADOR SOUTH
SAGUENAY
STRAIT OF BELLE ISLE
ST BARBE NORTH
WHITE BAY NORTH
GREY ISLANDS
GULF OF ST LAWRENCE
LONG RANGE MOUNTAINS
ST BARBE SOUTH
WHITE BAY SOUTH
NOTRE DAME BAY
GREEN BAY
HUMBER WEST
HUMBER EAST
CORNER BROOK
Limit of study area
PORT AU PORT
ST GEORGES
GRAND FALLS
GANDER
HERMITAGE
FORTUNE BAY
Lambert Conformal Conic Projection
Projection conique conforme de L
SCALE 1:1,000,000 ÉCHELLE
REFERENCE
LÉGENDE
INDEX TO BOUNDARIES
TABLEAU DES LIMITES
QUEBEC
NEWFOUNDLAND
ADJOINING MAPS
CARTES ADJACENTES
CARTWRIGHT
ANTICOSTI ISLAND
CORNER BROOK
HALIFAX
ST. JOHNS

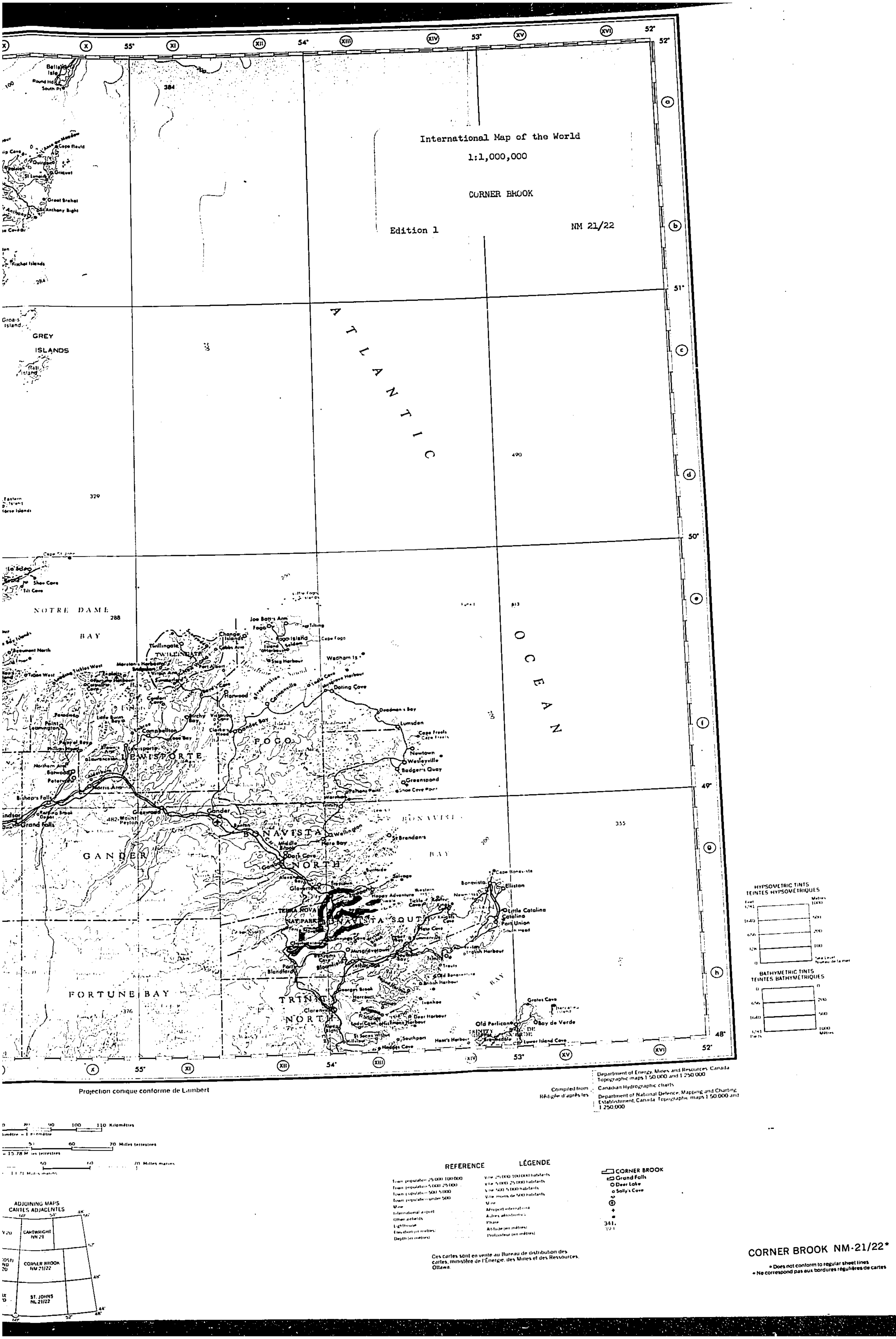

International Map of the World
1:1,000,000
CORNER BROOK
Edition 1
NM 21/22
ATLANTIC
OCEAN
GREY
ISLANDS
NOTRE DAME
BAY
LEWISPORTE
GANDER
BONAVISTA
NORTH
FORTUNE BAY
TRINITY
NORTH
Fogo Island
Twillingate
Gander
Grand Falls
Bishop's Falls
Botwood
Wesleyville
Badger's Quay
Greenspond
Lumsden
Newtown
Musgravetown
Clarenville
Bonavista
Elliston
Little Catalina
Catalina
Port Union
Bay de Verde
Old Perlican
Grates Cove
Lower Island Cove
Southport
Heart's Delight
55°
54°
53°
52°
51°
50°
49°
48°
Projection conique conforme de Lambert
Compiled from
Rédigée d'après les
Department of Energy, Mines and Resources, Canada
Topographic maps 1 50 000 and 1 250 000
Canadian Hydrographic charts
Department of National Defence, Mapping and Charting Establishment, Canada. Topographic maps 1 50 000 and 1 250 000
HYPSOMETRIC TINTS
TEINTES HYPSOMÉTRIQUES
BATHYMETRIC TINTS
TEINTES BATHYMÉTRIQUES
ADJOINING MAPS
CARTES ADJACENTES
CARTWRIGHT
NN 21
CORNER BROOK
NM 21/22
ST. JOHNS
NL 21/22
REFERENCE
LÉGENDE
CORNER BROOK
Grand Falls
Deer Lake
Sally's Cove
341.
Ces cartes sont en vente au Bureau de distribution des cartes, ministère de l'Énergie, des Mines et des Ressources, Ottawa.
CORNER BROOK NM-21/22*
* Does not conform to regular sheet lines
* Ne correspond pas aux bordures régulières de cartes

PLATE II

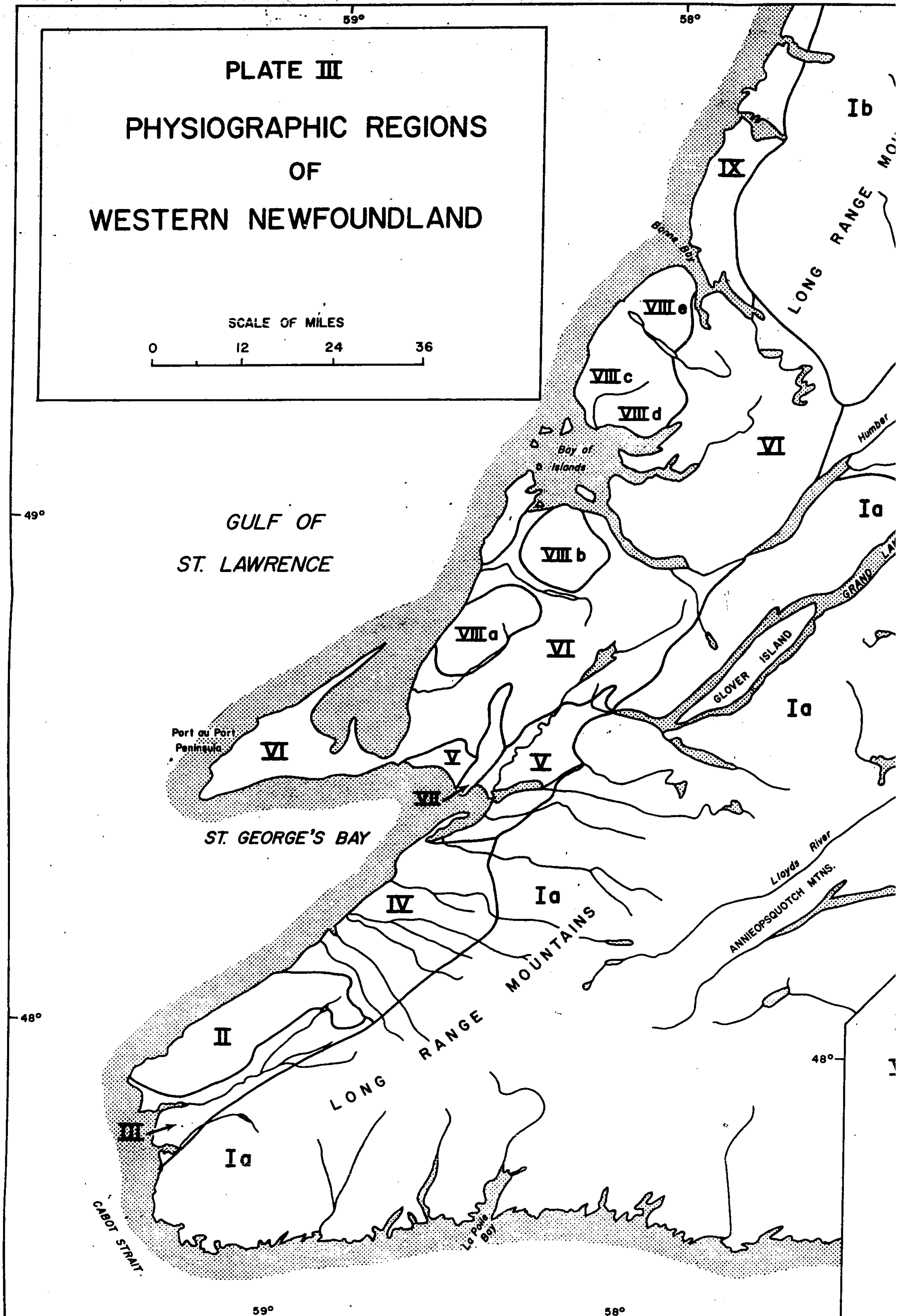
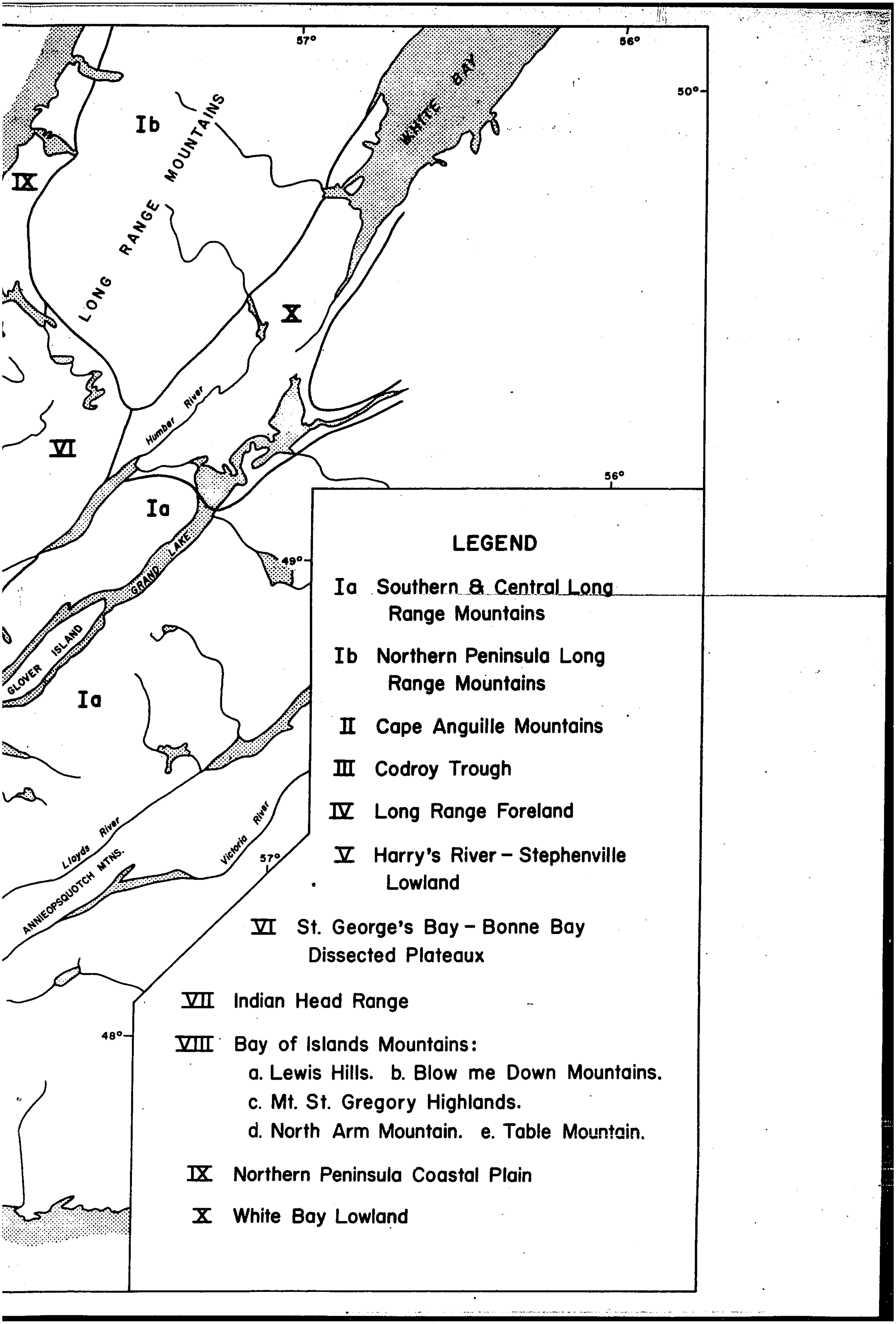


PLATE III
PHYSIOGRAPHIC REGIONS
OF
WESTERN NEWFOUNDLAND
SCALE OF MILES
0
12
24
36
59°
58°
49°
48°
GULF OF
ST. LAWRENCE
ST. GEORGE'S BAY
Port au Port
Peninsula
Bonne Bay
Bay of
Islands
Humber
GRAND LAKE
GLOVER ISLAND
Lloyds River
ANNIEOPSQUOTCH MTNS.
LONG RANGE MOUNTAINS
CABOT STRAIT
La Poile Bay
Ia
Ib
II
III
IV
V
VI
VII
VIII a
VIII b
VIII c
VIII d
VIII e
IX



57°
56°
50°
49°
48°
WHITE BAY
LONG RANGE MOUNTAINS
Ib
IX
X
VI
Ia
Humber River
GRAND LAKE
GLOVER ISLAND
Lloyds River
Victoria River
ANNIEOPSQUOTCH MTNS.
LEGEND
Ia Southern & Central Long Range Mountains
Ib Northern Peninsula Long Range Mountains
II Cape Anguille Mountains
III Codroy Trough
IV Long Range Foreland
V Harry's River – Stephenville Lowland
VI St. George's Bay – Bonne Bay Dissected Plateaux
VII Indian Head Range
VIII Bay of Islands Mountains:
a. Lewis Hills. b. Blow me Down Mountains.
c. Mt. St. Gregory Highlands.
d. North Arm Mountain. e. Table Mountain.
IX Northern Peninsula Coastal Plain
X White Bay Lowland

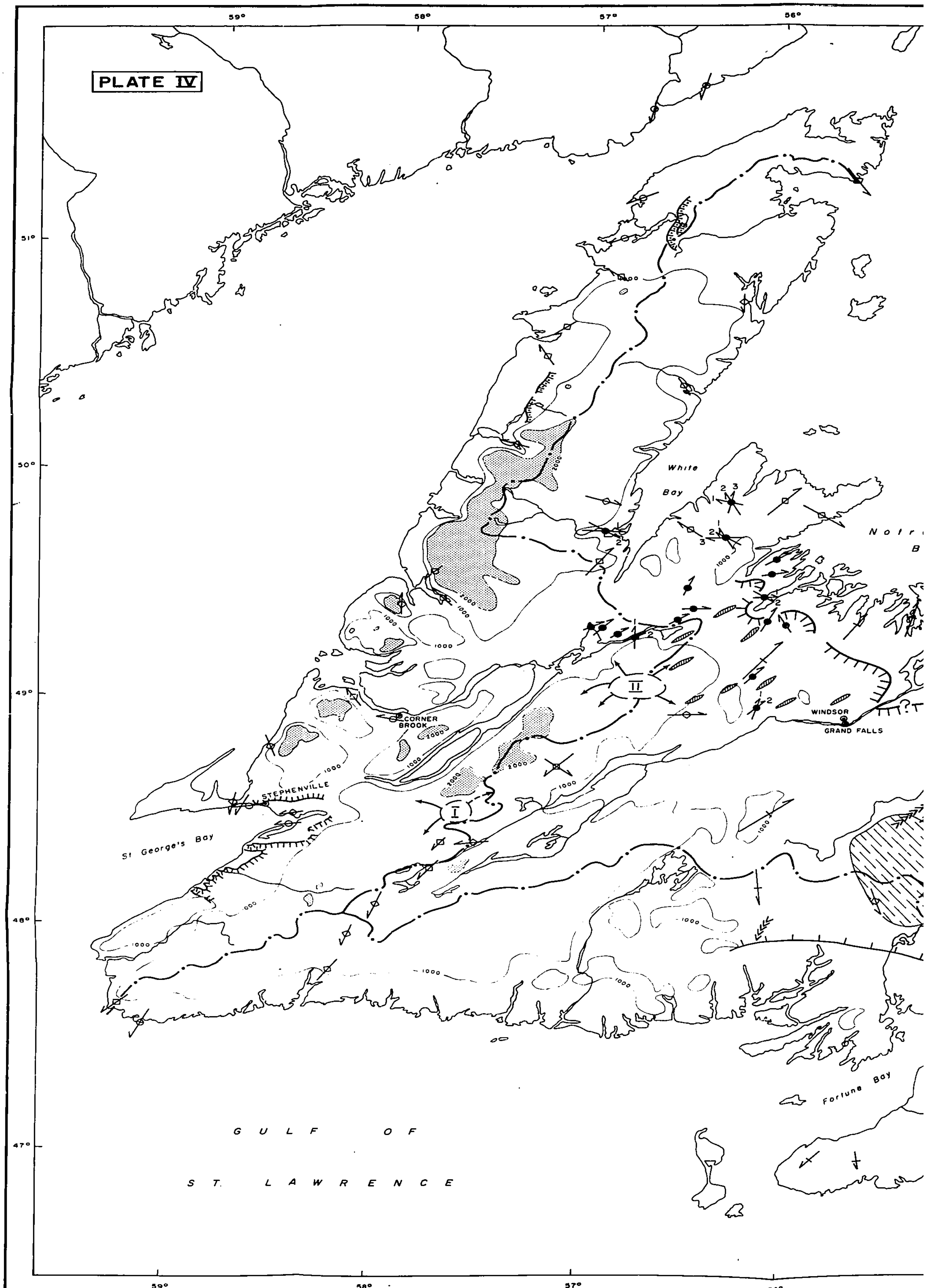
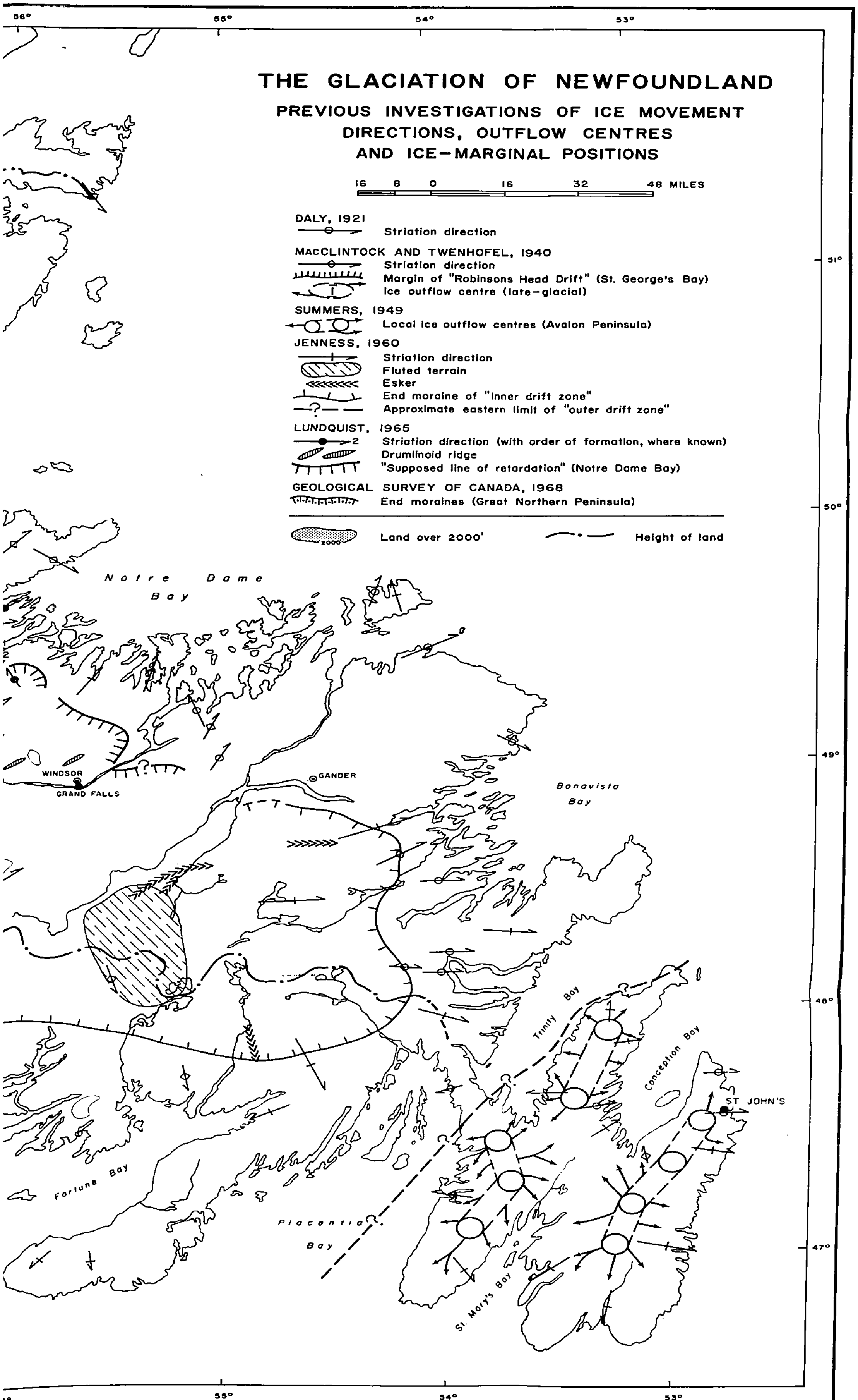


PLATE IV
59°
58°
57°
56°
51°
50°
49°
48°
47°
White Bay
Notr
B
CORNER BROOK
STEPHENVILLE
WINDSOR
GRAND FALLS
St George's Bay
Fortune Bay
GULF OF
ST. LAWRENCE
I
II
1000
2000



THE GLACIATION OF NEWFOUNDLAND
PREVIOUS INVESTIGATIONS OF ICE MOVEMENT DIRECTIONS, OUTFLOW CENTRES AND ICE-MARGINAL POSITIONS
16 8 0 16 32 48 MILES
DALY, 1921
Striation direction
MACCLINTOCK AND TWENHOFEL, 1940
Striation direction
Margin of "Robinsons Head Drift" (St. George's Bay)
Ice outflow centre (late-glacial)
SUMMERS, 1949
Local ice outflow centres (Avalon Peninsula)
JENNESS, 1960
Striation direction
Fluted terrain
Esker
End moraine of "inner drift zone"
Approximate eastern limit of "outer drift zone"
LUNDQUIST, 1965
Striation direction (with order of formation, where known)
Drumlinoid ridge
"Supposed line of retardation" (Notre Dame Bay)
GEOLOGICAL SURVEY OF CANADA, 1968
End moraines (Great Northern Peninsula)
Land over 2000'
Height of land
Notre Dame Bay
WINDSOR
GRAND FALLS
GANDER
Bonavista Bay
Trinity Bay
Conception Bay
ST JOHN'S
Fortune Bay
Placentia Bay
St Mary's Bay
55°
54°
53°
51°
50°
49°
48°
47°

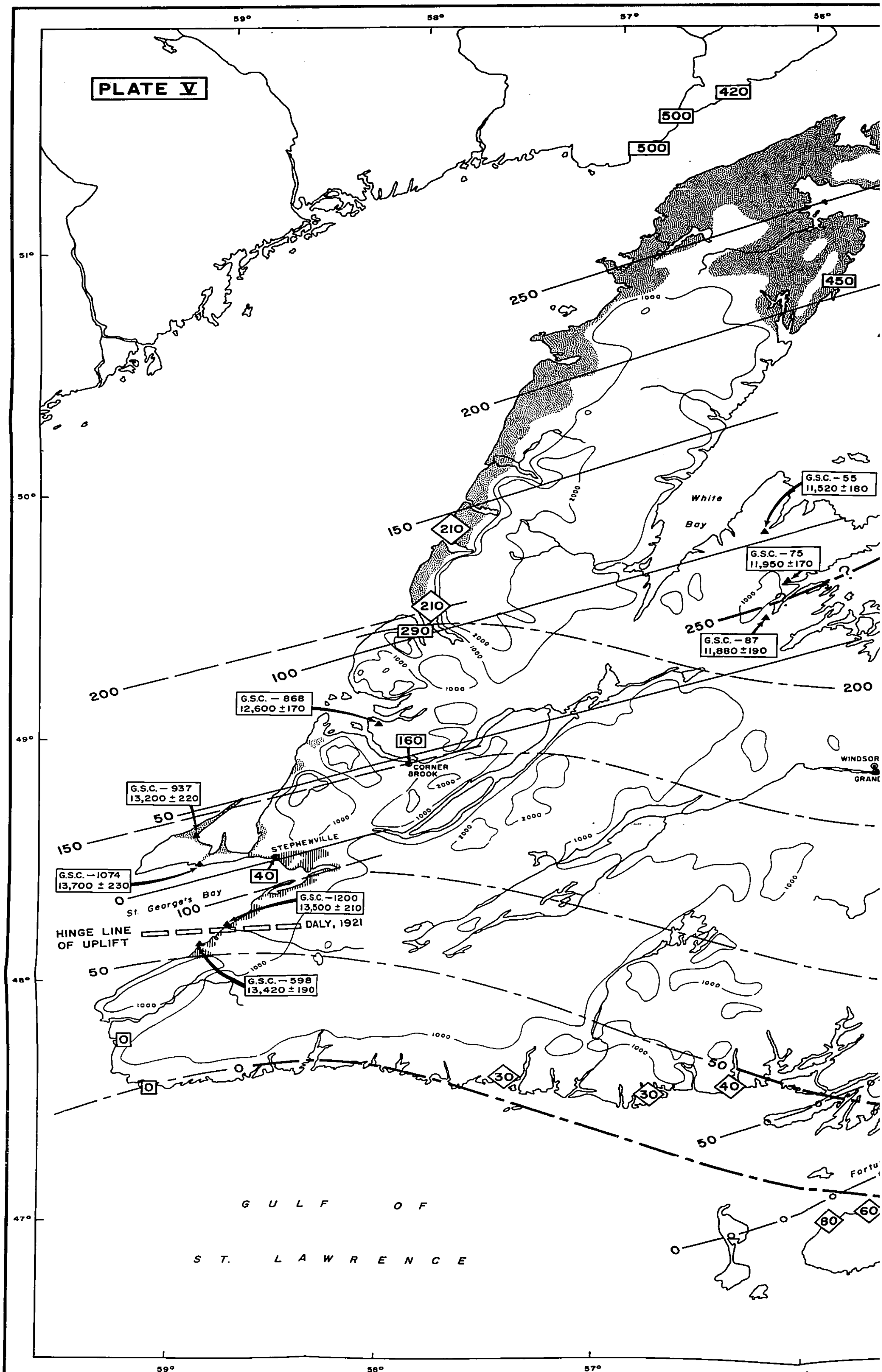


PLATE V
GULF OF ST. LAWRENCE
White Bay
St. George's Bay
HINGE LINE OF UPLIFT
DALY, 1921
CORNER BROOK
STEPHENVILLE
WINDSOR
G.S.C. – 55
11,520 ± 180
G.S.C. – 75
11,950 ± 170
G.S.C. – 87
11,880 ± 190
G.S.C. – 868
12,600 ± 170
G.S.C. – 937
13,200 ± 220
G.S.C. – 1074
13,700 ± 230
G.S.C. – 1200
13,500 ± 210
G.S.C. – 598
13,420 ± 190

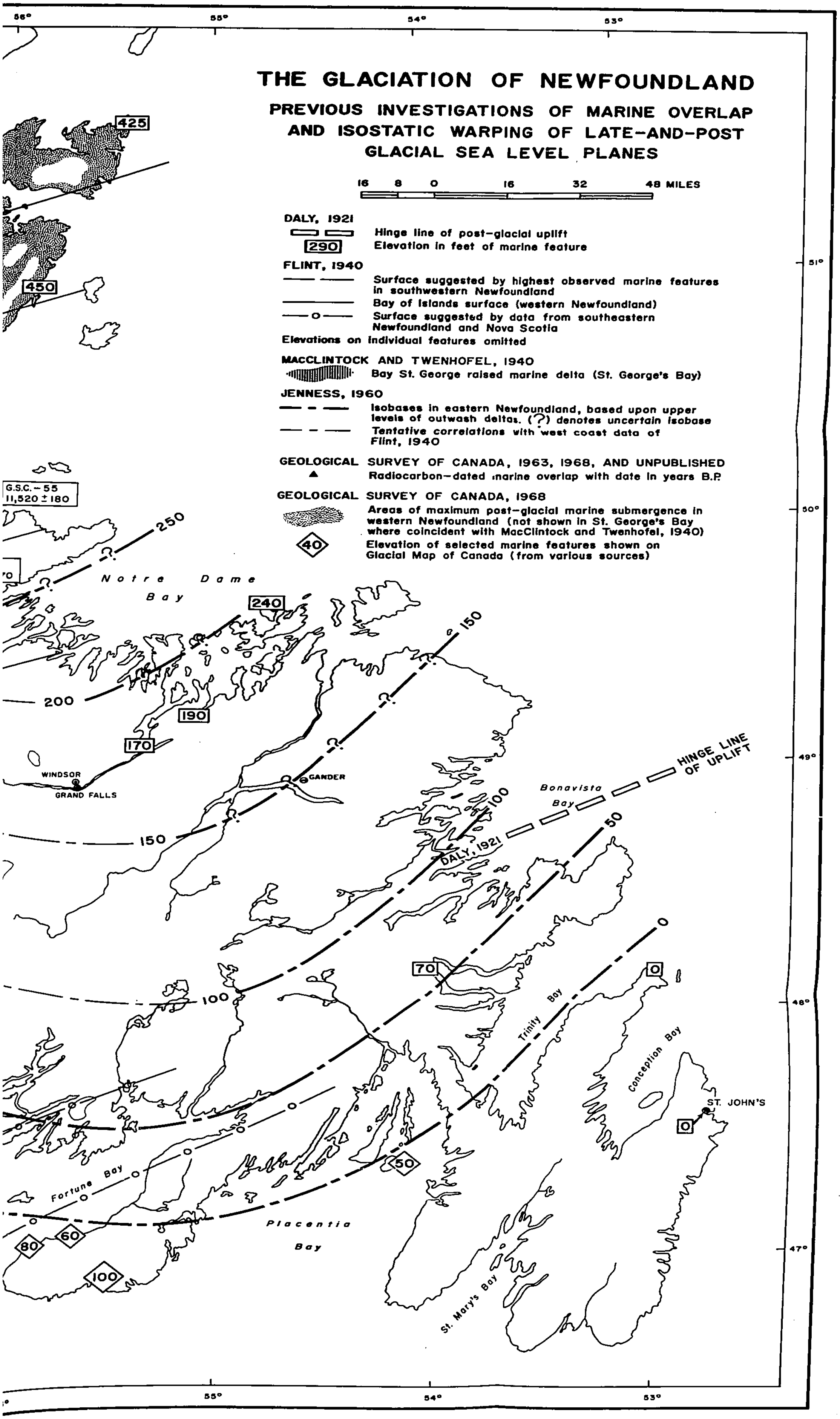
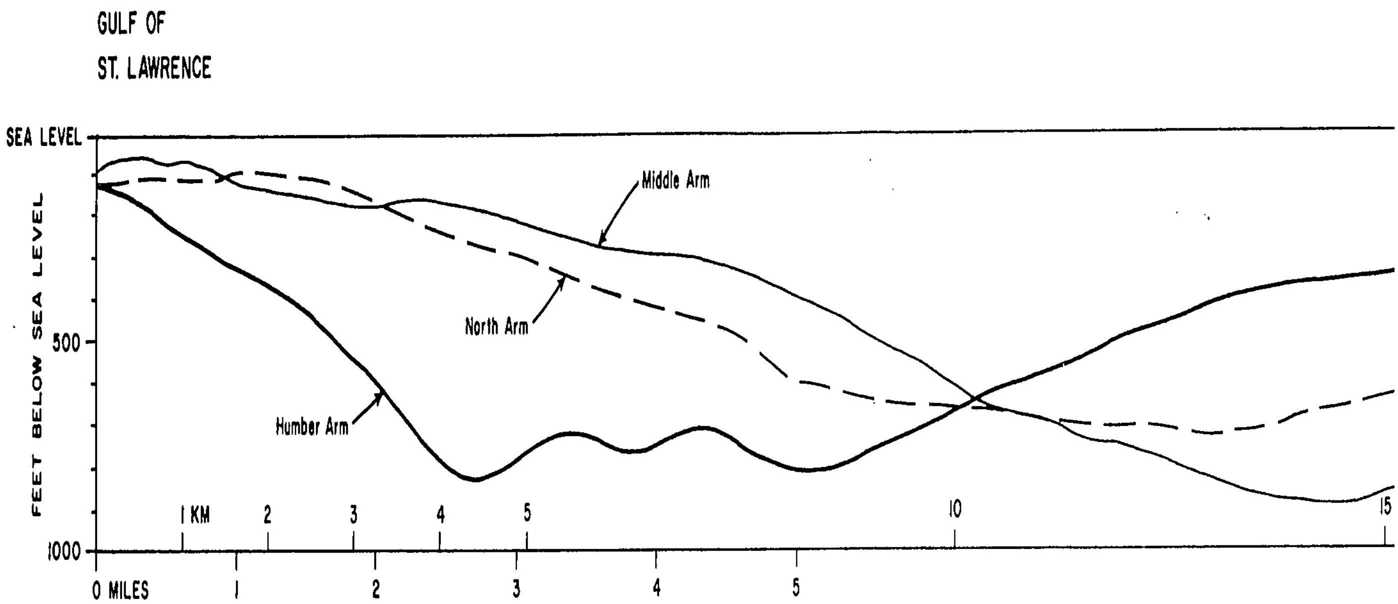
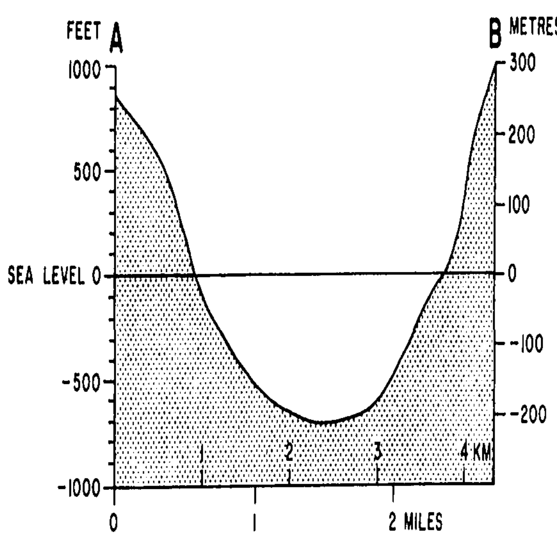
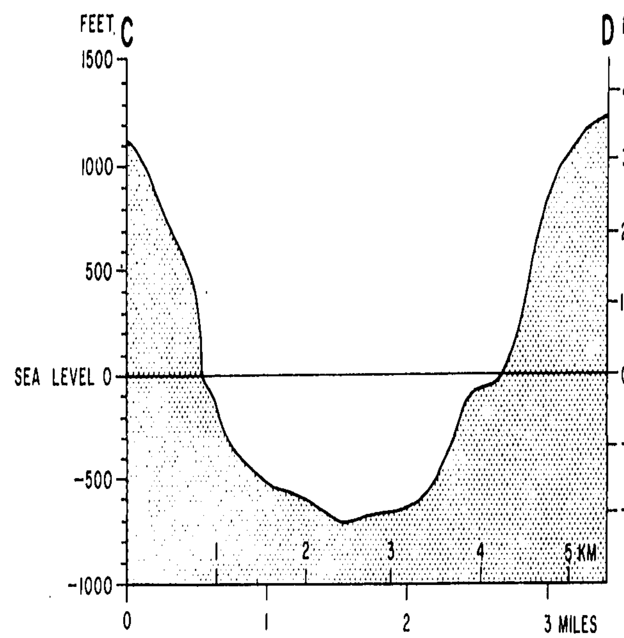

THE GLACIATION OF NEWFOUNDLAND
PREVIOUS INVESTIGATIONS OF MARINE OVERLAP AND ISOSTATIC WARPING OF LATE-AND-POST GLACIAL SEA LEVEL PLANES
16 8 0 16 32 48 MILES
DALY, 1921
Hinge line of post-glacial uplift
290
Elevation in feet of marine feature
FLINT, 1940
Surface suggested by highest observed marine features in southwestern Newfoundland
Bay of Islands surface (western Newfoundland)
Surface suggested by data from southeastern Newfoundland and Nova Scotia
Elevations on individual features omitted
MACCLINTOCK AND TWENHOFEL, 1940
Bay St. George raised marine delta (St. George's Bay)
JENNESS, 1960
Isobases in eastern Newfoundland, based upon upper levels of outwash deltas. (?) denotes uncertain isobase
Tentative correlations with west coast data of Flint, 1940
GEOLOGICAL SURVEY OF CANADA, 1963, 1968, AND UNPUBLISHED
Radiocarbon-dated marine overlap with date in years B.P.
GEOLOGICAL SURVEY OF CANADA, 1968
Areas of maximum post-glacial marine submergence in western Newfoundland (not shown in St. George's Bay where coincident with MacClintock and Twenhofel, 1940)
40
Elevation of selected marine features shown on Glacial Map of Canada (from various sources)
56°
55°
54°
53°
51°
50°
49°
48°
47°
425
450
G.S.C.-55
11,520 ± 180
250
Notre Dame Bay
240
150
200
190
170
WINDSOR
GRAND FALLS
GANDER
HINGE LINE OF UPLIFT
Bonavista Bay
100
DALY, 1921
50
150
0
70
100
Trinity Bay
Conception Bay
ST. JOHN'S
0
Fortune Bay
50
Placentia Bay
80
60
100
St. Mary's Bay

PLATE VI



Vertical Exaggeration ≃ x 8

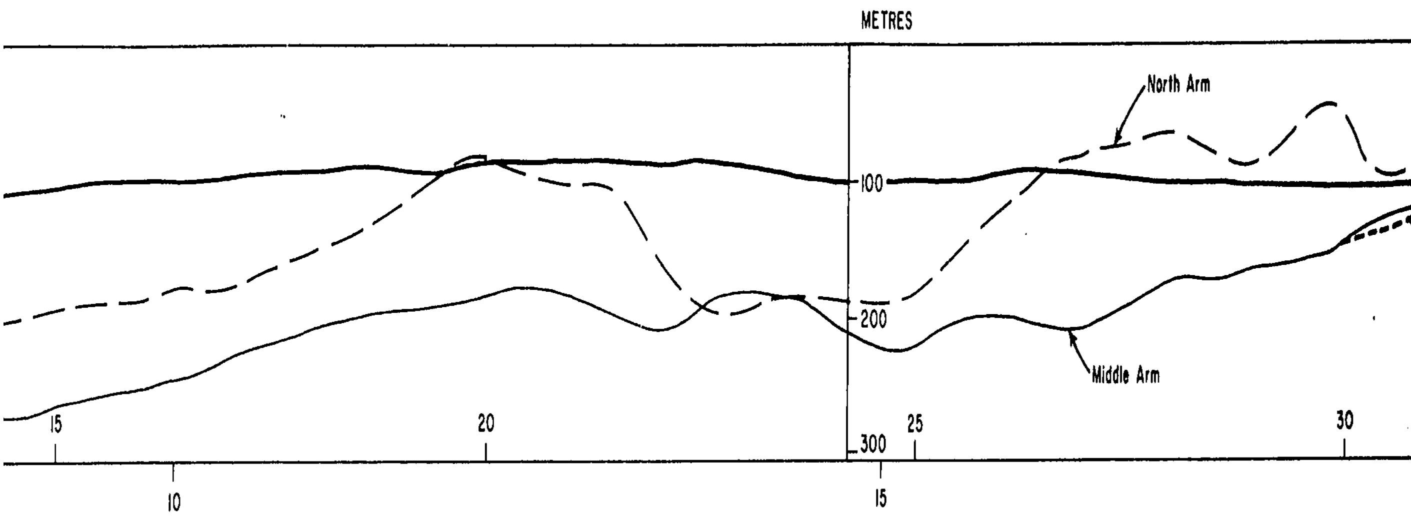


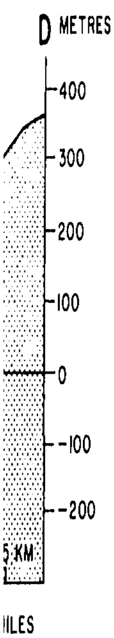


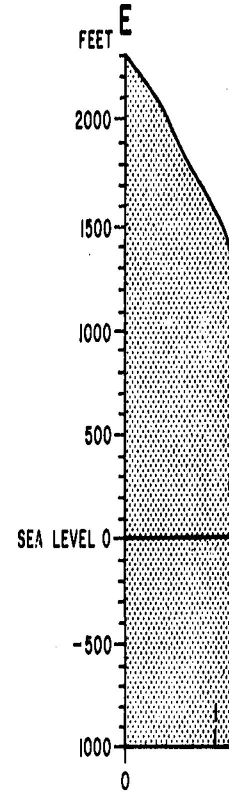
Vertical Exaggeration ≃ x 4

PROFILES OF FJORDS OF THE BAY OF ISLANDS

(For location of long and cross profiles, see figure 14)

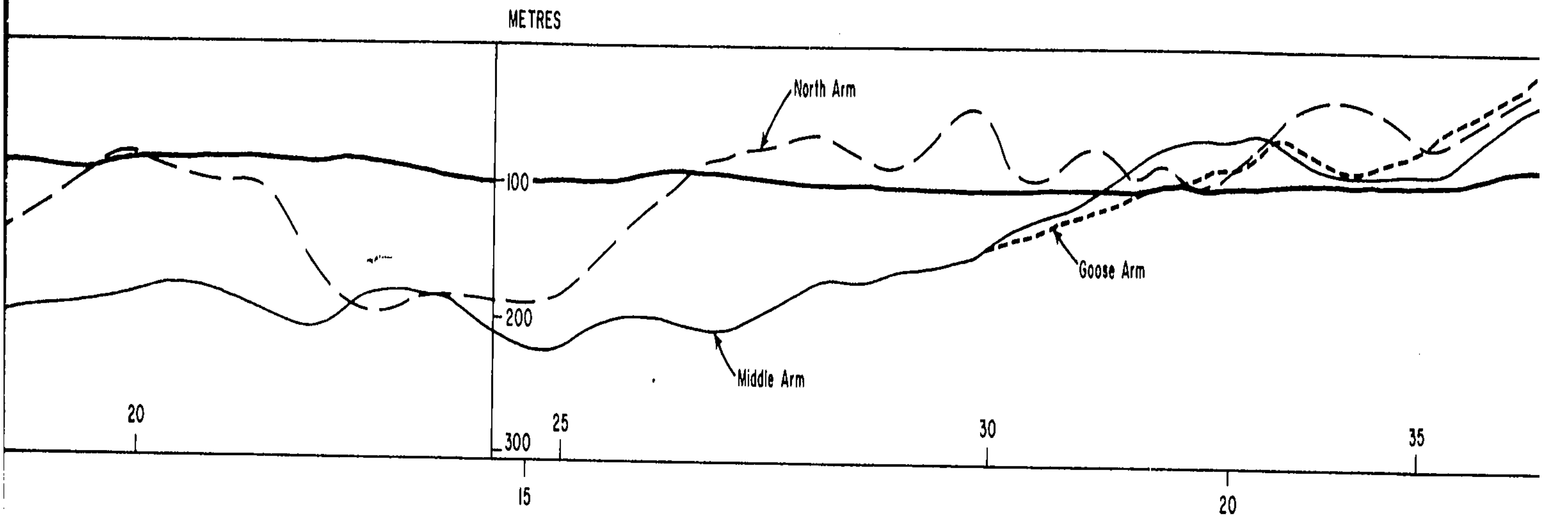


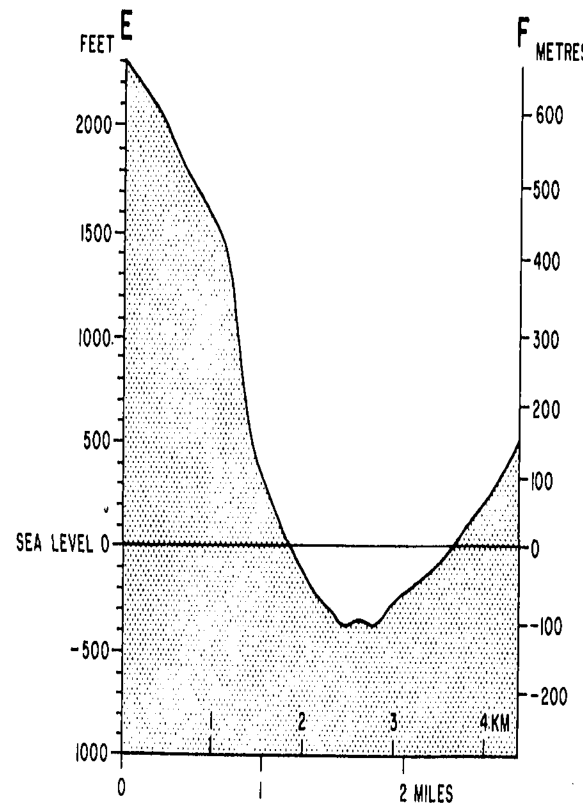


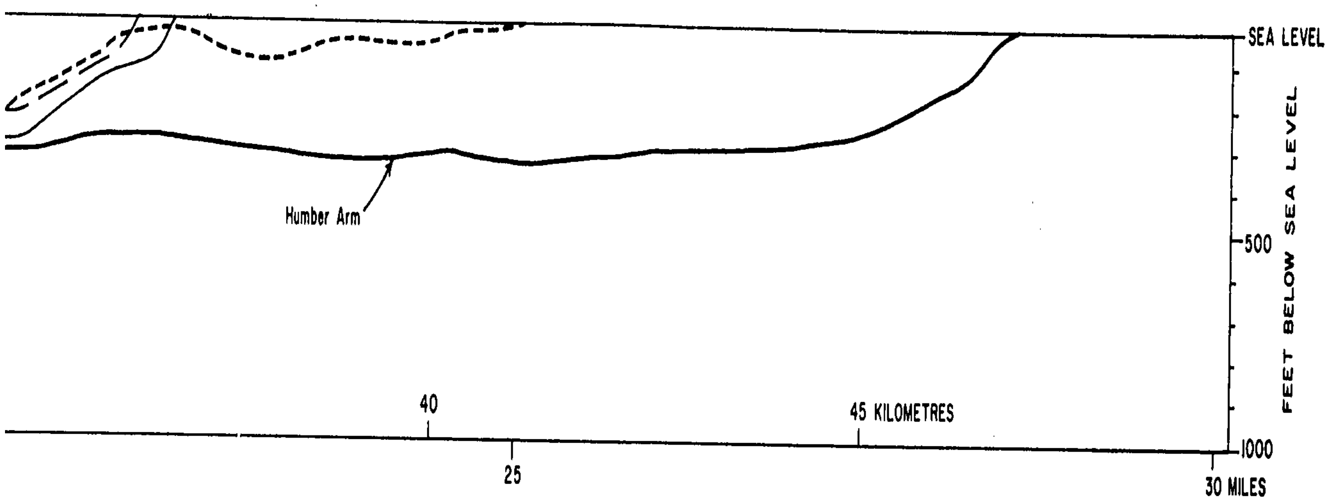


PROFILES OF FJORDS OF THE BAY OF ISLANDS

(For location of long and cross profiles, see figure 14)







SEA LEVEL
FEET BELOW SEA LEVEL
500
1000
Humber Arm
40
45 KILOMETRES
25
30 MILES

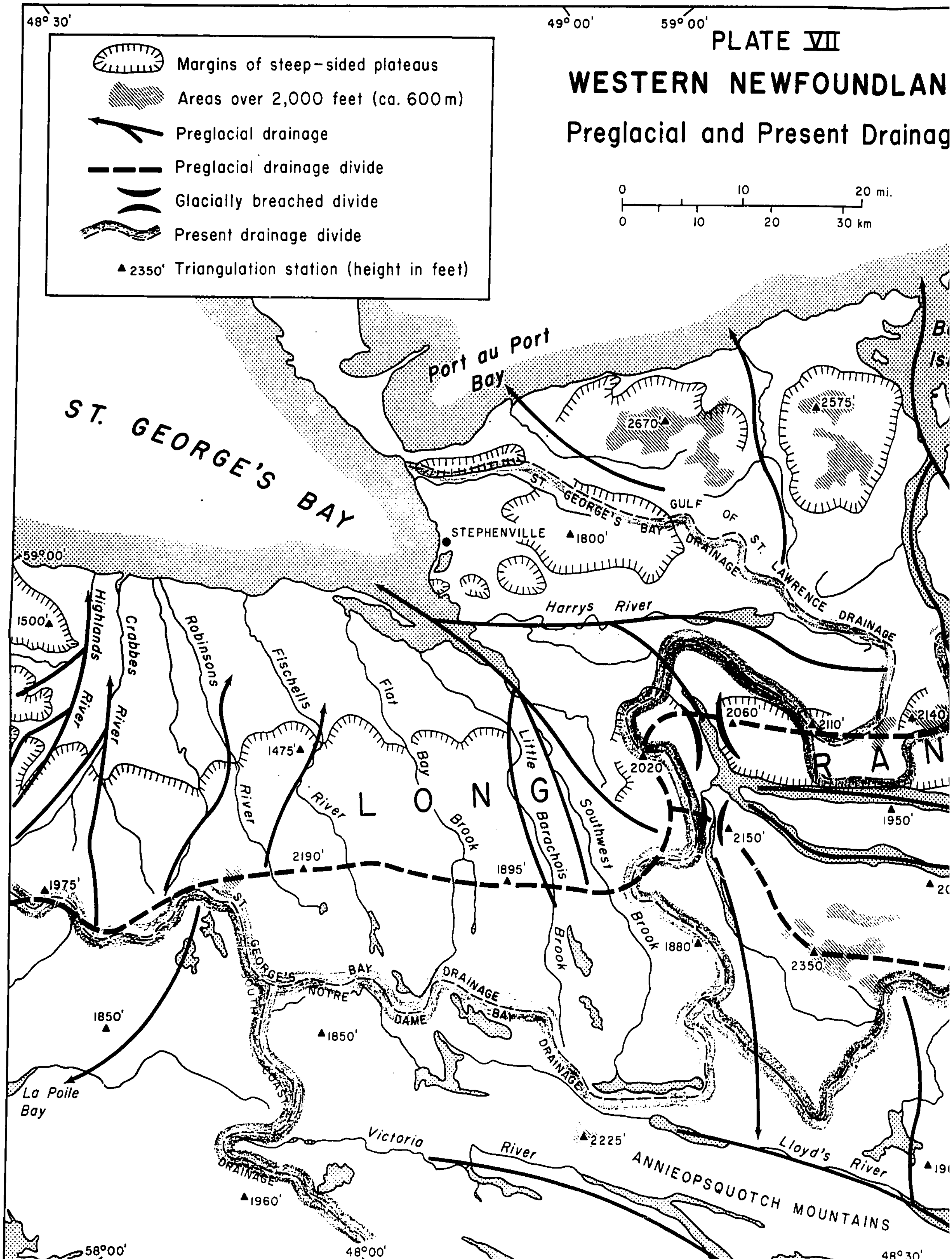
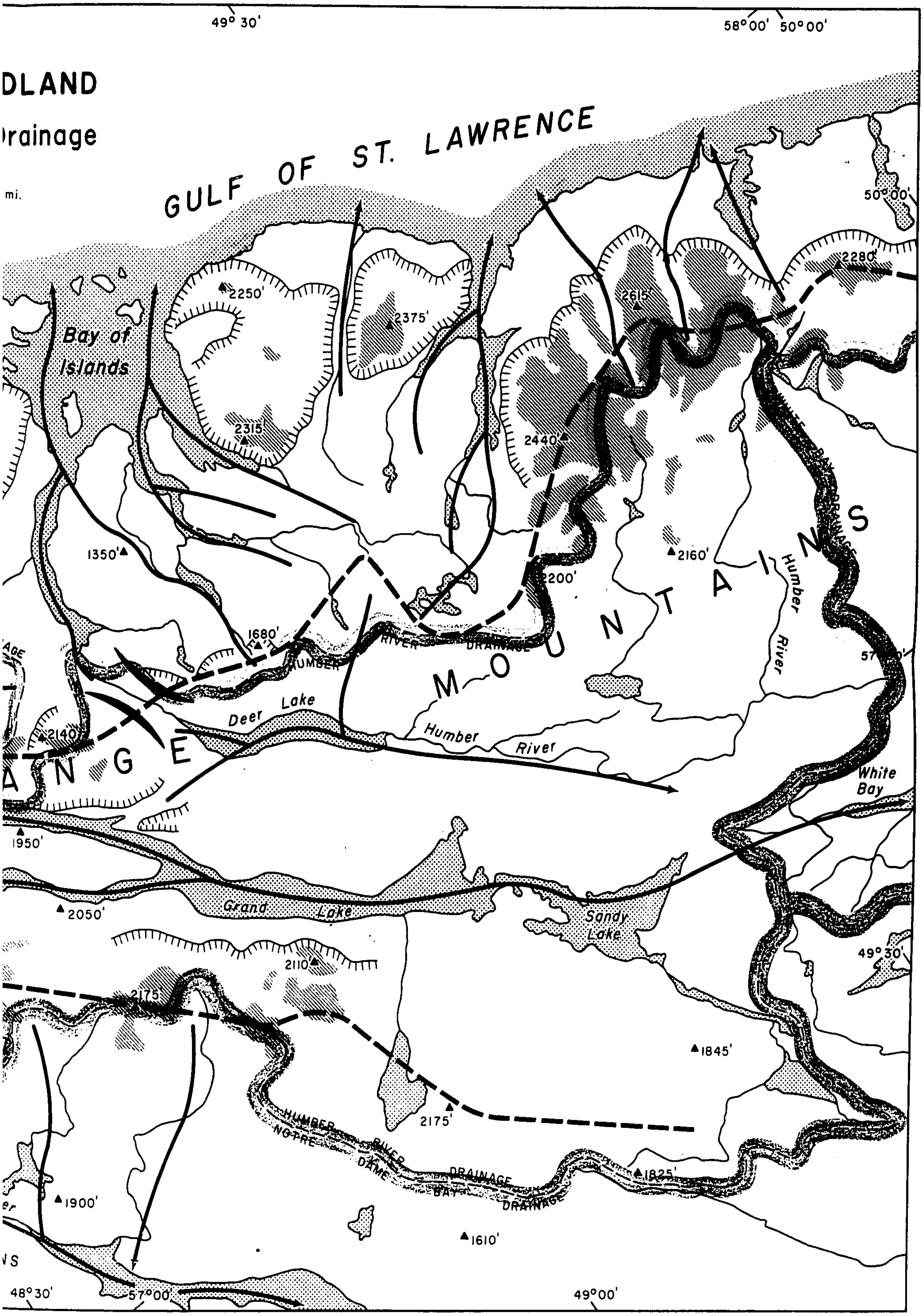
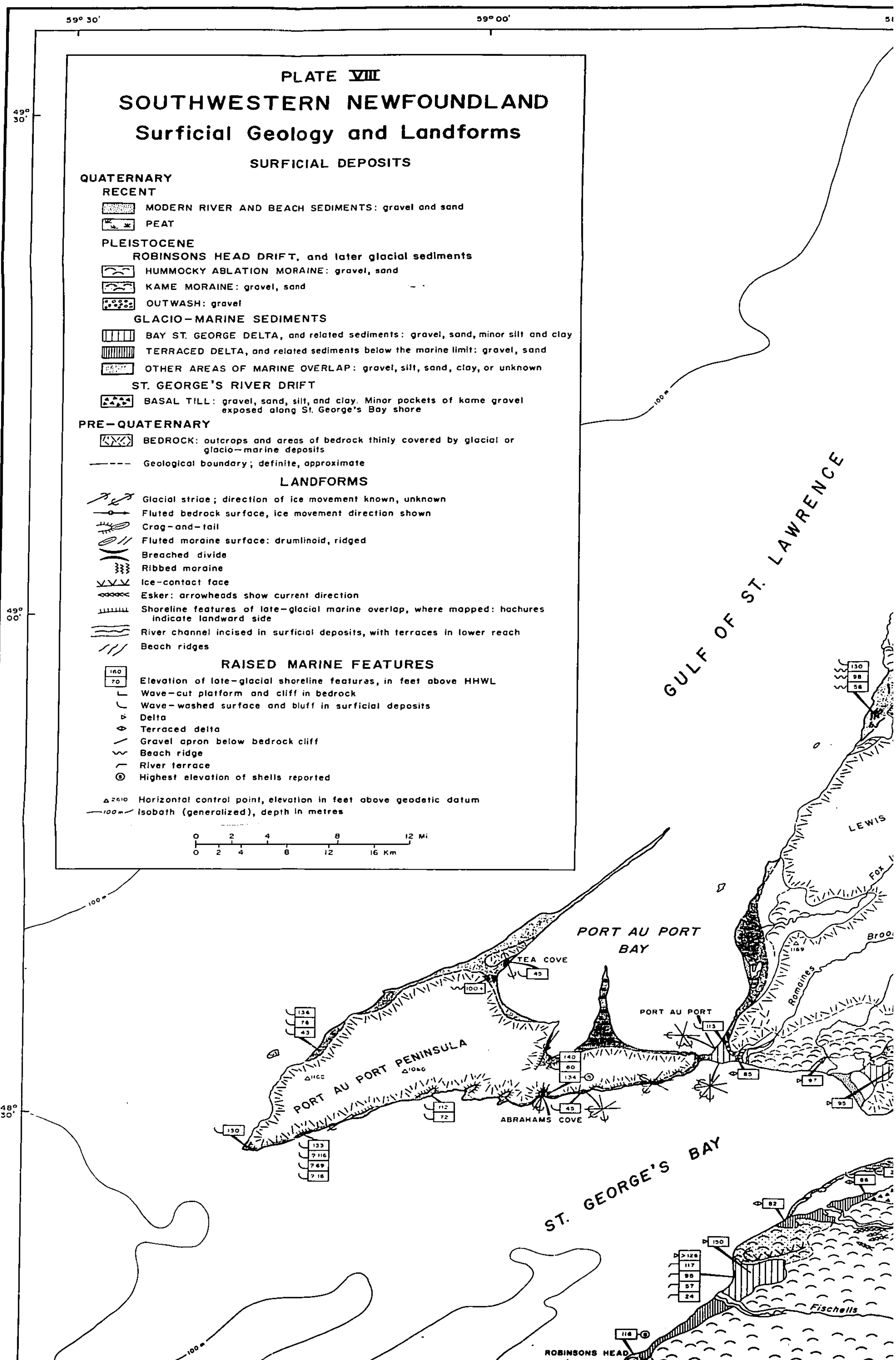
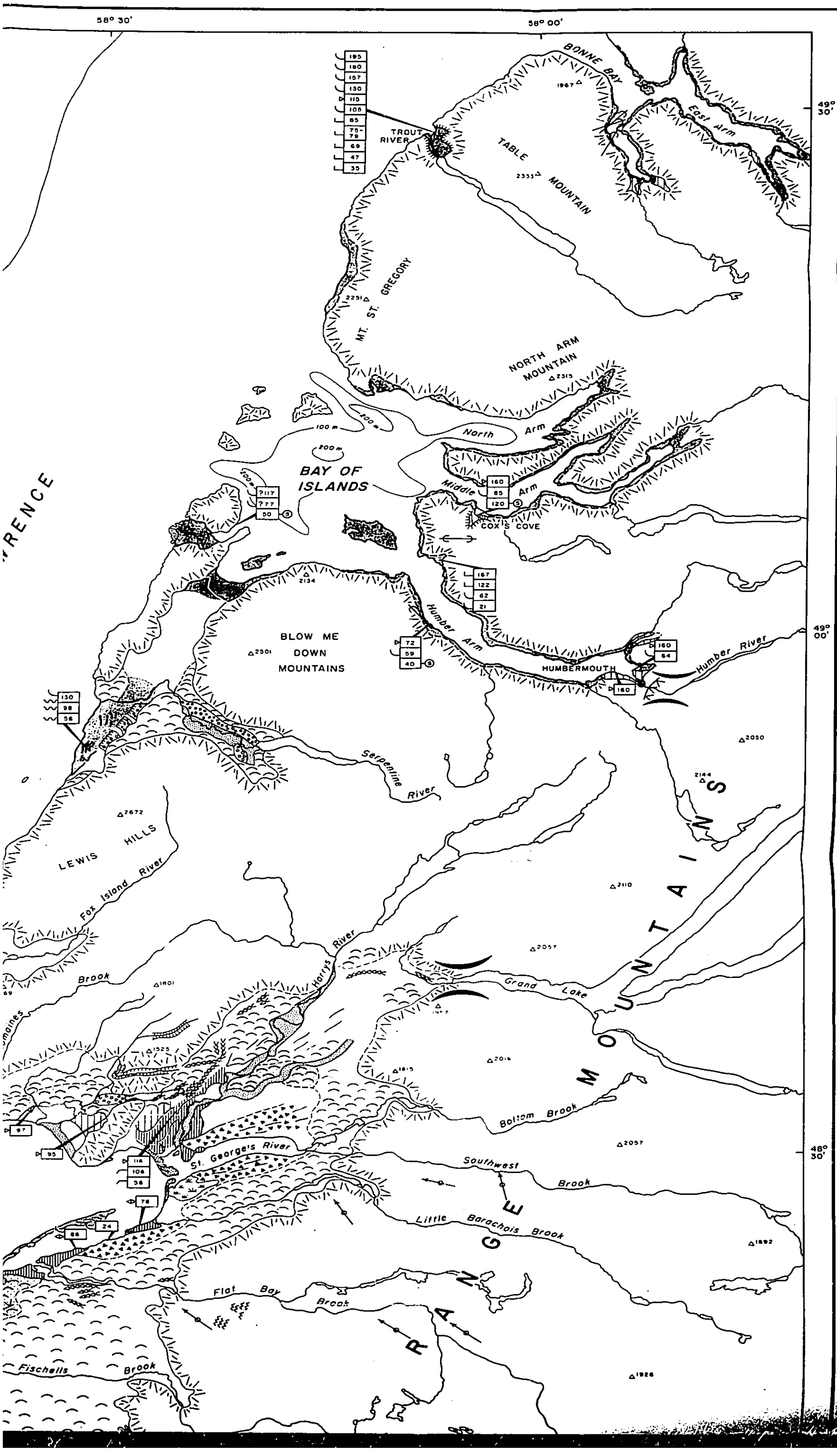


PLATE VII
WESTERN NEWFOUNDLAN
Preglacial and Present Drainag
Margins of steep-sided plateaus
Areas over 2,000 feet (ca. 600 m)
Preglacial drainage
Preglacial drainage divide
Glacially breached divide
Present drainage divide
2350' Triangulation station (height in feet)
0 10 20 mi.
0 10 20 30 km
48° 30'
49° 00'
59° 00'
ST. GEORGE'S BAY
Port au Port Bay
STEPHENVILLE
GULF OF ST. LAWRENCE DRAINAGE
ST. GEORGE'S BAY DRAINAGE
Harrys River
Highlands River
Crabbes River
Robinsons River
Fischells River
Flat Bay Brook
Little Barachois Brook
Southwest Brook
L O N G
R A N
ST. GEORGE'S BAY DRAINAGE
NOTRE DAME BAY DRAINAGE
SOUTH COAST DRAINAGE
La Poile Bay
Victoria River
Lloyd's River
ANNIEOPSQUOTCH MOUNTAINS
2670'
2575'
1800'
1500'
1475'
2060'
2110'
2140
2020'
1950'
2150'
2190'
1895'
1975'
1880'
2350'
1850'
1850'
2225'
1960'
58°00'
48°00'
48°30'

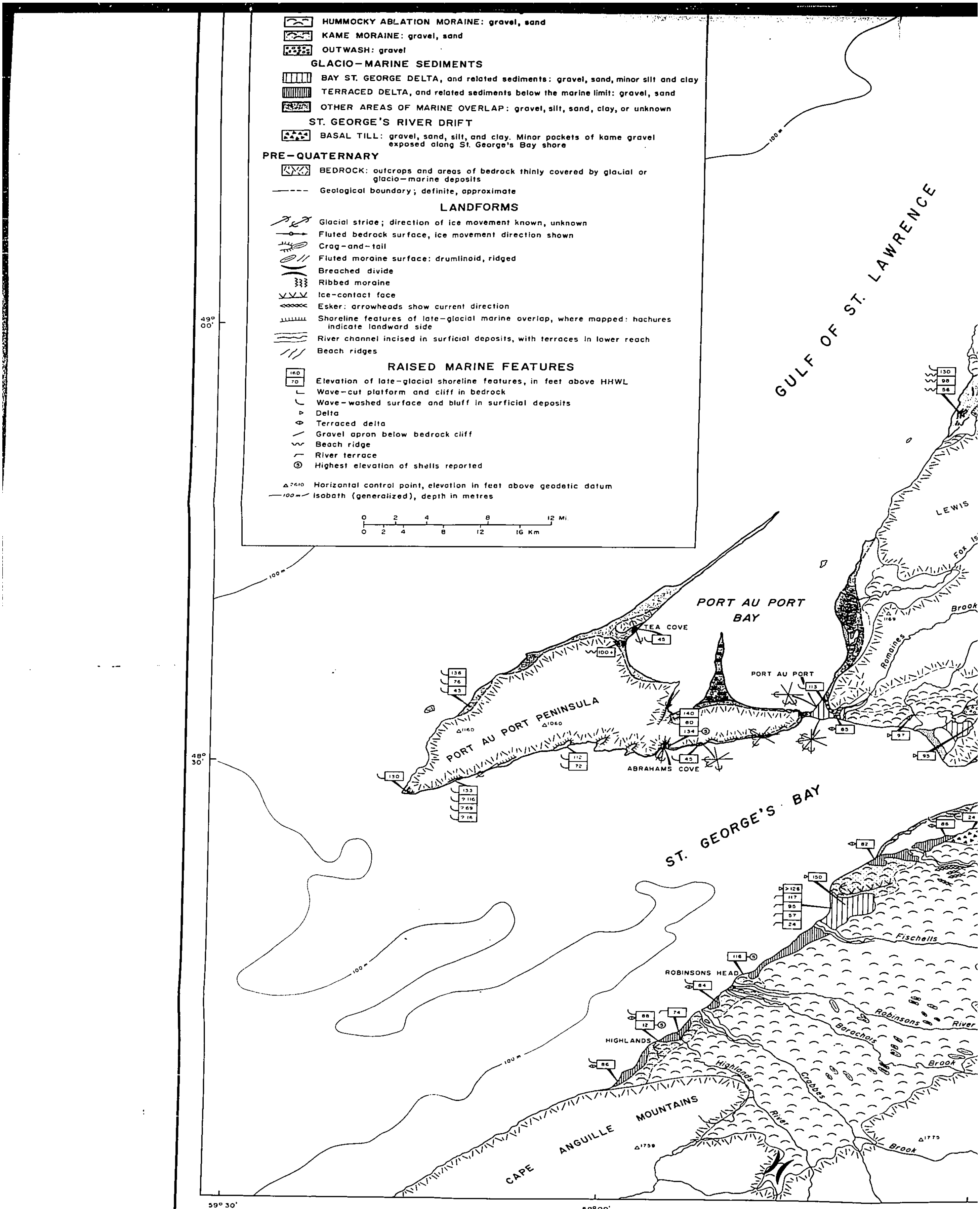


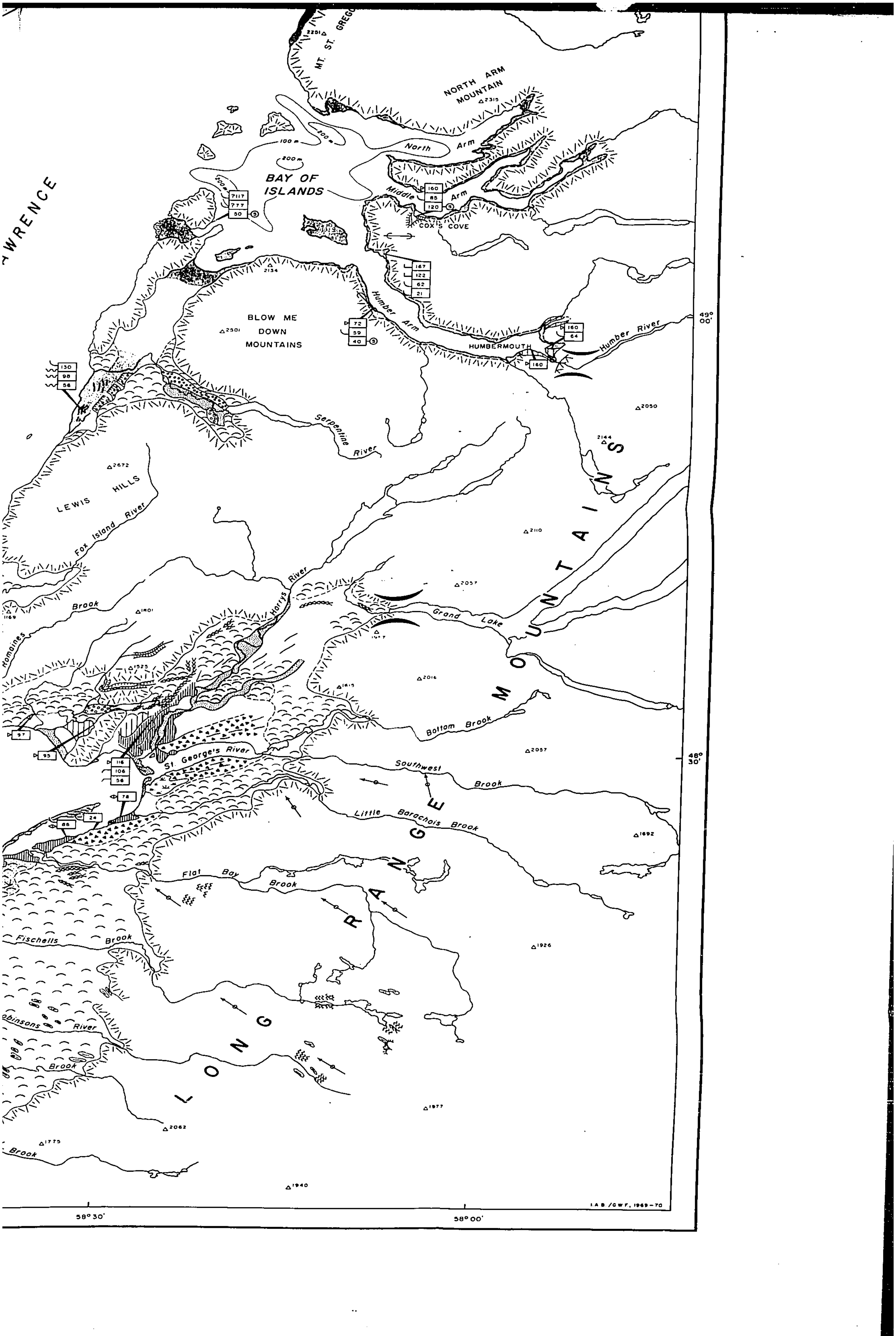
DLAND
rainage
mi.
49° 30'
58°00'
50°00'
GULF OF ST. LAWRENCE
Bay of Islands
2250'
2375'
2315'
2615'
2280'
2440'
2160'
2200'
1350'
1680'
2140'
1950'
2050'
2110'
2175'
2175'
1845'
1825'
1900'
1610'
MOUNTAINS
ANGE
Humber River
Deer Lake
Humber River
HUMBER RIVER DRAINAGE
Grand Lake
Sandy Lake
White Bay
HUMBER RIVER DRAINAGE
NOTRE DAME BAY DRAINAGE
49°30'
48°30'
57°00'
49°00'

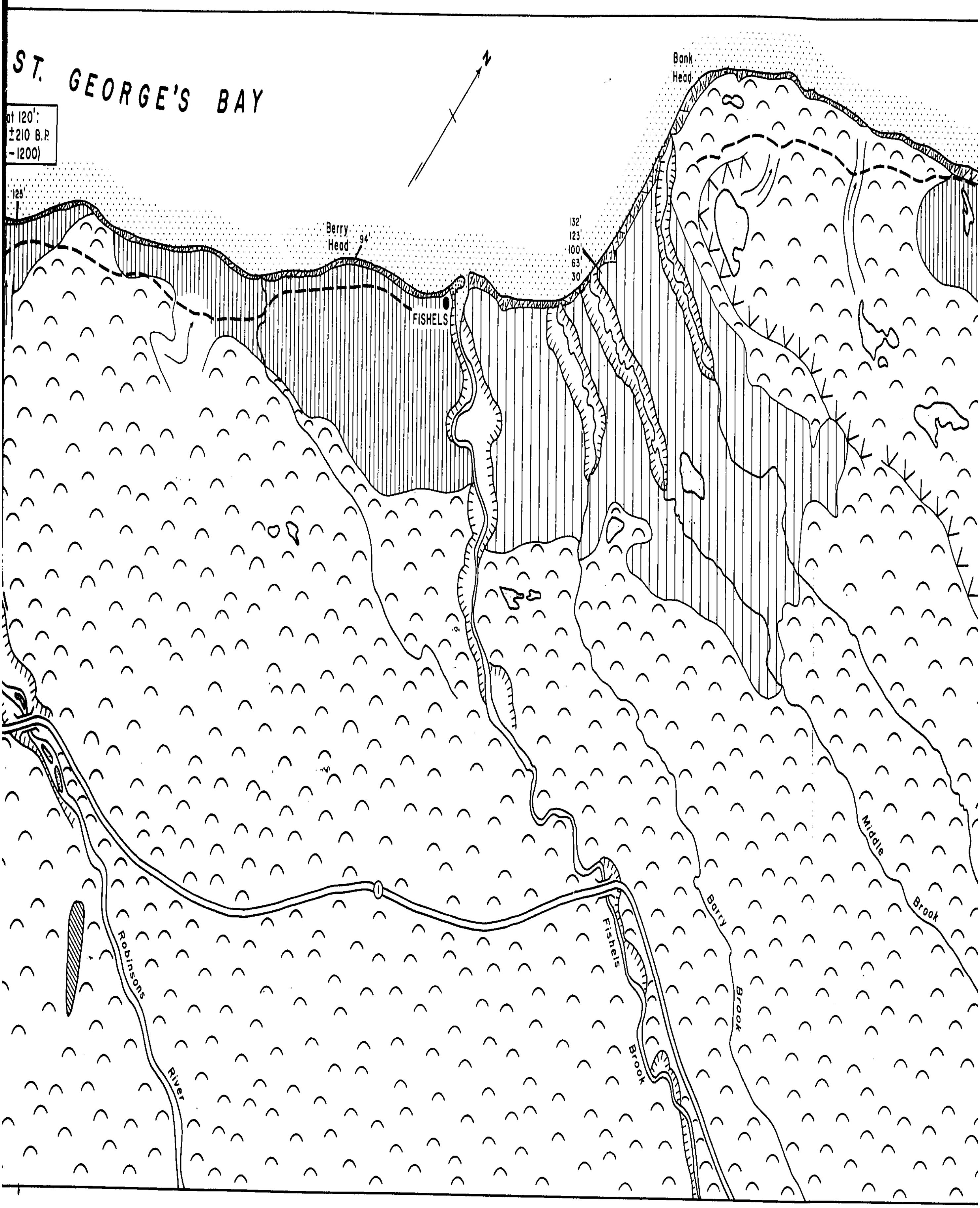

59° 30'
59° 00'
49° 30'
49° 00'
48° 30'
PLATE VIII
SOUTHWESTERN NEWFOUNDLAND
Surficial Geology and Landforms
SURFICIAL DEPOSITS
QUATERNARY
RECENT
MODERN RIVER AND BEACH SEDIMENTS: gravel and sand
PEAT
PLEISTOCENE
ROBINSONS HEAD DRIFT, and later glacial sediments
HUMMOCKY ABLATION MORAINE: gravel, sand
KAME MORAINE: gravel, sand
OUTWASH: gravel
GLACIO-MARINE SEDIMENTS
BAY ST. GEORGE DELTA, and related sediments: gravel, sand, minor silt and clay
TERRACED DELTA, and related sediments below the marine limit: gravel, sand
OTHER AREAS OF MARINE OVERLAP: gravel, silt, sand, clay, or unknown
ST. GEORGE'S RIVER DRIFT
BASAL TILL: gravel, sand, silt, and clay. Minor pockets of kame gravel exposed along St. George's Bay shore
PRE-QUATERNARY
BEDROCK: outcrops and areas of bedrock thinly covered by glacial or glacio-marine deposits
Geological boundary; definite, approximate
LANDFORMS
Glacial striae; direction of ice movement known, unknown
Fluted bedrock surface, ice movement direction shown
Crag-and-tail
Fluted moraine surface: drumlinoid, ridged
Breached divide
Ribbed moraine
Ice-contact face
Esker: arrowheads show current direction
Shoreline features of late-glacial marine overlap, where mapped: hachures indicate landward side
River channel incised in surficial deposits, with terraces in lower reach
Beach ridges
RAISED MARINE FEATURES
Elevation of late-glacial shoreline features, in feet above HHWL
Wave-cut platform and cliff in bedrock
Wave-washed surface and bluff in surficial deposits
Delta
Terraced delta
Gravel apron below bedrock cliff
Beach ridge
River terrace
Highest elevation of shells reported
Horizontal control point, elevation in feet above geodetic datum
Isobath (generalized), depth in metres
0 2 4 8 12 Mi
0 2 4 8 12 16 Km
GULF OF ST. LAWRENCE
PORT AU PORT BAY
TEA COVE
PORT AU PORT
PORT AU PORT PENINSULA
ABRAHAMS COVE
ST. GEORGE'S BAY
ROBINSONS HEAD
LEWIS
Romaines
Fischells



58° 30'
58° 00'
49° 30'
49° 00'
48° 30'
BONNE BAY
East Arm
TROUT RIVER
TABLE MOUNTAIN
MT. ST. GREGORY
NORTH ARM MOUNTAIN
North Arm
BAY OF ISLANDS
Middle Arm
COX'S COVE
Humber Arm
BLOW ME DOWN MOUNTAINS
HUMBERMOUTH
Humber River
Serpentine River
LEWIS HILLS
Fox Island River
Harrys River
Grand Lake
Brook
Bottom Brook
St. George's River
Southwest Brook
Little Barachois Brook
Flat Bay Brook
Fischells Brook
RANGE
MOUNTAINS
RENCE


HUMMOCKY ABLATION MORAINE: gravel, sand
KAME MORAINE: gravel, sand
OUTWASH: gravel
GLACIO-MARINE SEDIMENTS
BAY ST. GEORGE DELTA, and related sediments: gravel, sand, minor silt and clay
TERRACED DELTA, and related sediments below the marine limit: gravel, sand
OTHER AREAS OF MARINE OVERLAP: gravel, silt, sand, clay, or unknown
ST. GEORGE'S RIVER DRIFT
BASAL TILL: gravel, sand, silt, and clay. Minor pockets of kame gravel exposed along St. George's Bay shore
PRE-QUATERNARY
BEDROCK: outcrops and areas of bedrock thinly covered by glacial or glacio-marine deposits
Geological boundary; definite, approximate
LANDFORMS
Glacial striae; direction of ice movement known, unknown
Fluted bedrock surface, ice movement direction shown
Crag-and-tail
Fluted moraine surface: drumlinoid, ridged
Breached divide
Ribbed moraine
Ice-contact face
Esker: arrowheads show current direction
Shoreline features of late-glacial marine overlap, where mapped: hachures indicate landward side
River channel incised in surficial deposits, with terraces in lower reach
Beach ridges
RAISED MARINE FEATURES
160
70
Elevation of late-glacial shoreline features, in feet above HHWL
Wave-cut platform and cliff in bedrock
Wave-washed surface and bluff in surficial deposits
Delta
Terraced delta
Gravel apron below bedrock cliff
Beach ridge
River terrace
Highest elevation of shells reported
Horizontal control point, elevation in feet above geodetic datum
Isobath (generalized), depth in metres
0 2 4 8 12 Mi.
0 2 4 8 12 16 Km
GULF OF ST. LAWRENCE
PORT AU PORT BAY
TEA COVE
PORT AU PORT
PORT AU PORT PENINSULA
ABRAHAMS COVE
ST. GEORGE'S BAY
LEWIS
Fox Is
Brook
Romaines
Fischells
ROBINSONS HEAD
Robinsons River
Barachois Brook
HIGHLANDS
Highlands River
Crabbes River
Brook
CAPE ANGUILLE MOUNTAINS
100 m
49° 00'
48° 30'
59° 30'
59° 00'
58° 30'


NORTH ARM MOUNTAIN
MT. ST. GREG
North Arm
Middle Arm
BAY OF ISLANDS
COX'S COVE
Humber Arm
BLOW ME DOWN MOUNTAINS
HUMBERMOUTH
Humber River
Serpentine River
LEWIS HILLS
Fox Island River
Harrys River
Grand Lake
Brook
Bottom Brook
St. George's River
Southwest Brook
Little Barachois Brook
Flat Bay Brook
Fischells Brook
Robinsons River
LONG RANGE MOUNTAINS
49° 00'
48° 30'
58° 30'
58° 00'
I.A.B./G.W.F., 1969-70

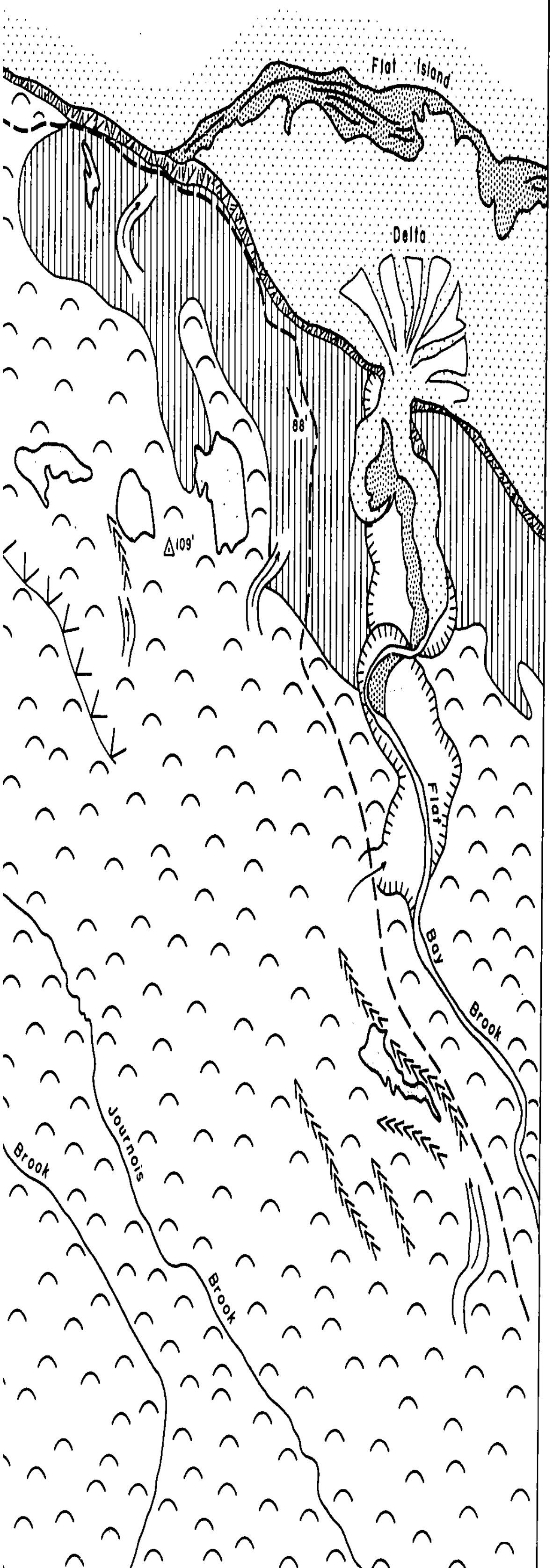


ST. GEORGE'S BAY
N
at 120':
±210 B.P.
-1200)
125'
Berry
Head
94'
FISHELS
132'
123'
100'
63'
30'
Bank
Head
Robinsons
River
Fishels
Brook
Barry
Brook
Middle
Brook

PLATE IX

THE HARBOUR HEAD-BANK HEAD SECTION

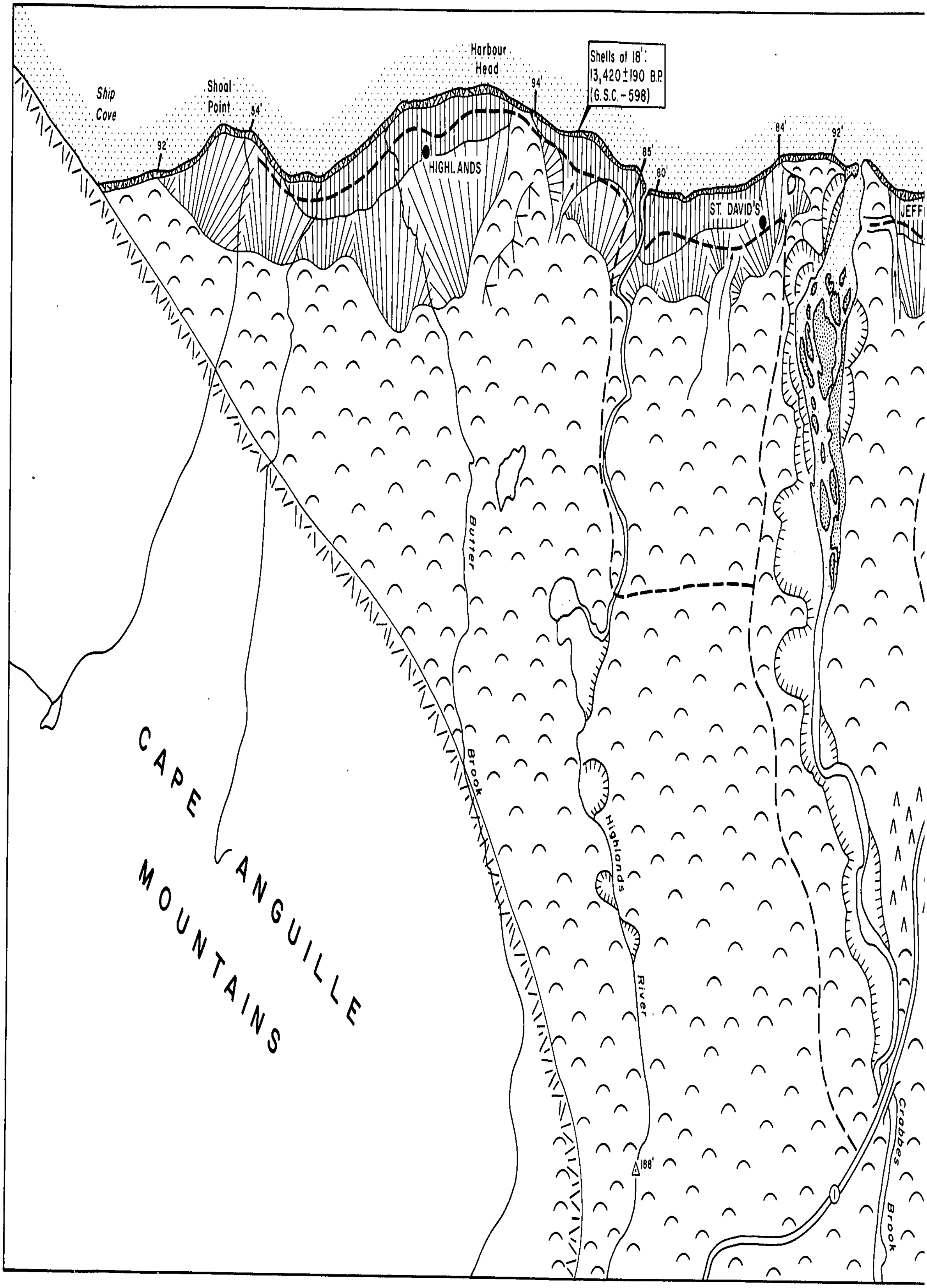
ST. GEORGE'S BAY



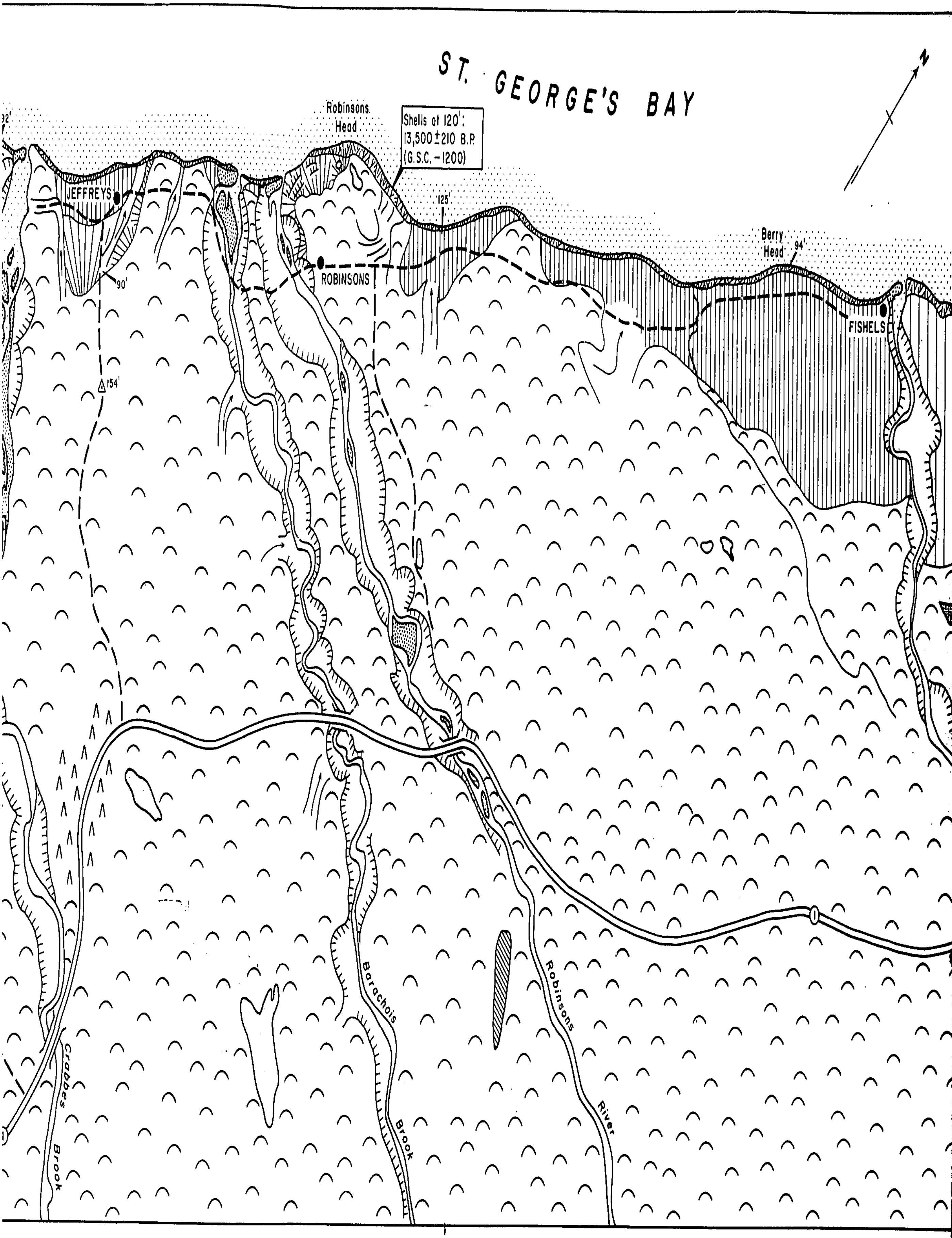
LEGEND

- Margin of bedrock
- Smooth hummocky Robinsons Head Drift
- Ice-contact face in end moraine
- Irregular, hummocky Robinsons Head Drift
- Kettle holes, water-filled
- Eskers in Robinsons Head Drift
- Transitional apron of outwash between Robinsons Head Drift end moraine and terraced Bay St. George Delta
- Bay St. George Delta (not affected by terracing)
- Terraced Bay St. George Delta surface (elevations indicated)
- Meltwater channel in Robinsons Head Drift
- Incised valley with terraces
- Coastal and riverine gravel bars
- Beach ridges on Flat Island gravel spit
- Coastal cliffs
- Drumlinoid ridge
- 188' Bench mark
- 90' Elevation above mean tide level, surveyed in this study
- Trans-Canada Highway
- Minor roads

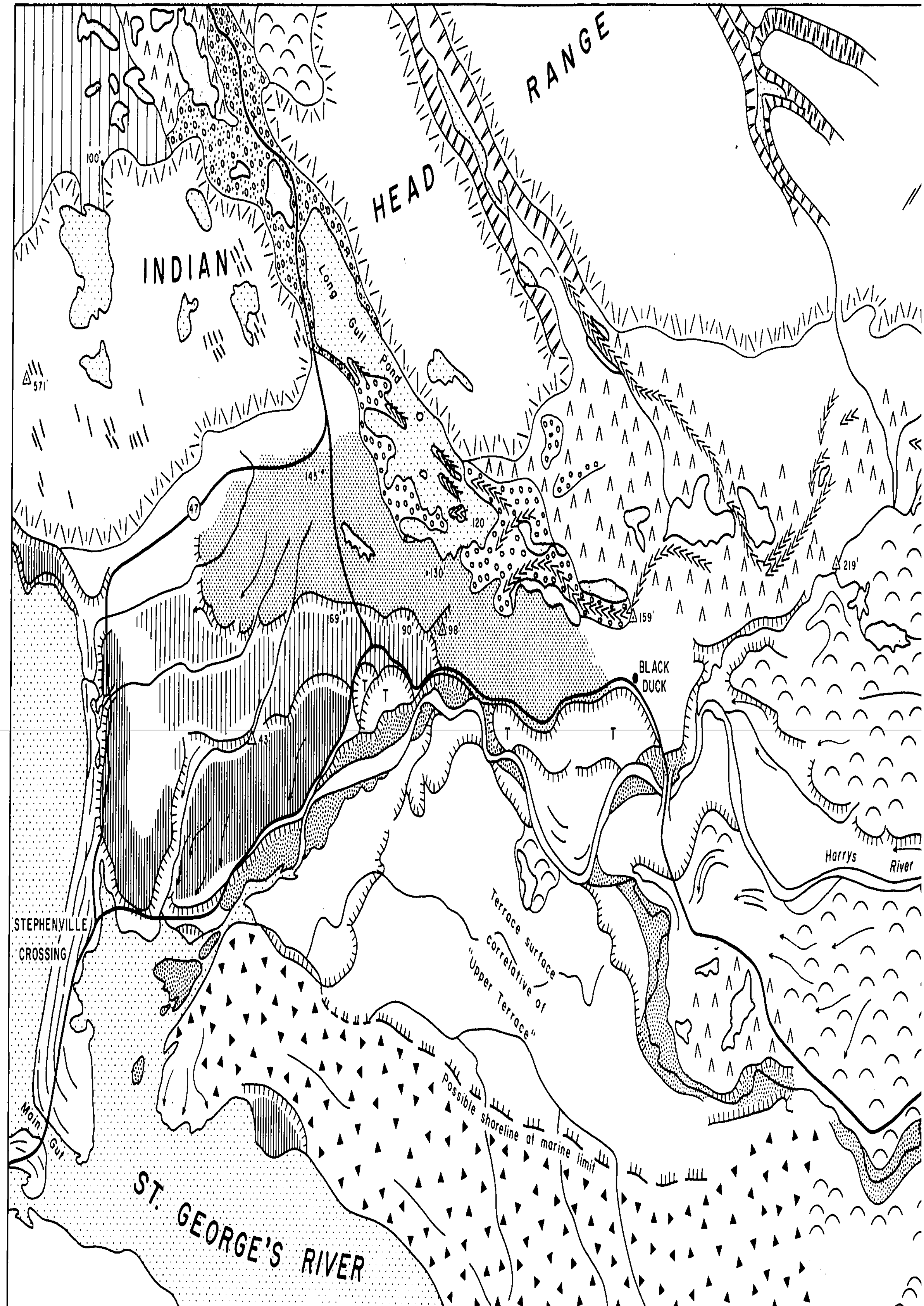
I.A.B./G.W.F.,1969-70



Ship Cove
Shoal Point
Harbour Head
Shells at 18':
13,420±190 B.P.
(G.S.C.-598)
92'
54'
94'
85'
80'
84'
92'
HIGHLANDS
ST. DAVID'S
JEFF
Butter Brook
Highlands River
CAPE ANGUILLE MOUNTAINS
188'
Crabbes Brook
1



ST. GEORGE'S BAY
N
Robinsons Head
Shells at 120'; 13,500±210 B.P. (G.S.C.-1200)
JEFFREYS
90'
125'
ROBINSONS
Berry Head
94'
FISHELS
154'
Barachois Brook
Robinsons River
Crabbes Brook


RANGE
HEAD
INDIAN
Long Gull Pond
571'
100'
145'
47
120
130
169'
90'
98
159'
219'
BLACK DUCK
43
Harrys River
Terrace surface correlative of "Upper Terrace"
STEPHENVILLE CROSSING
Main Gut
Possible shoreline at marine limit
ST. GEORGE'S RIVER

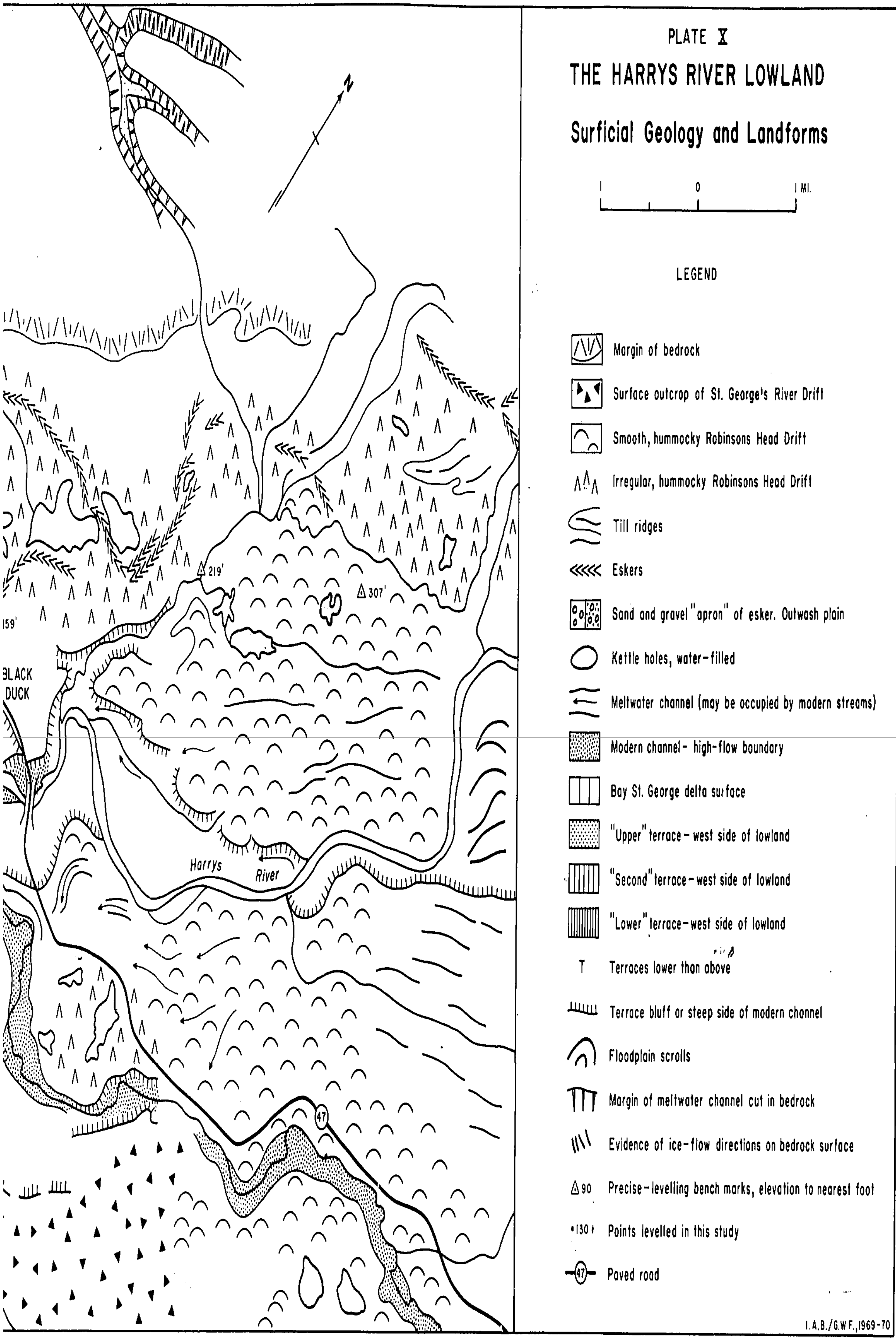
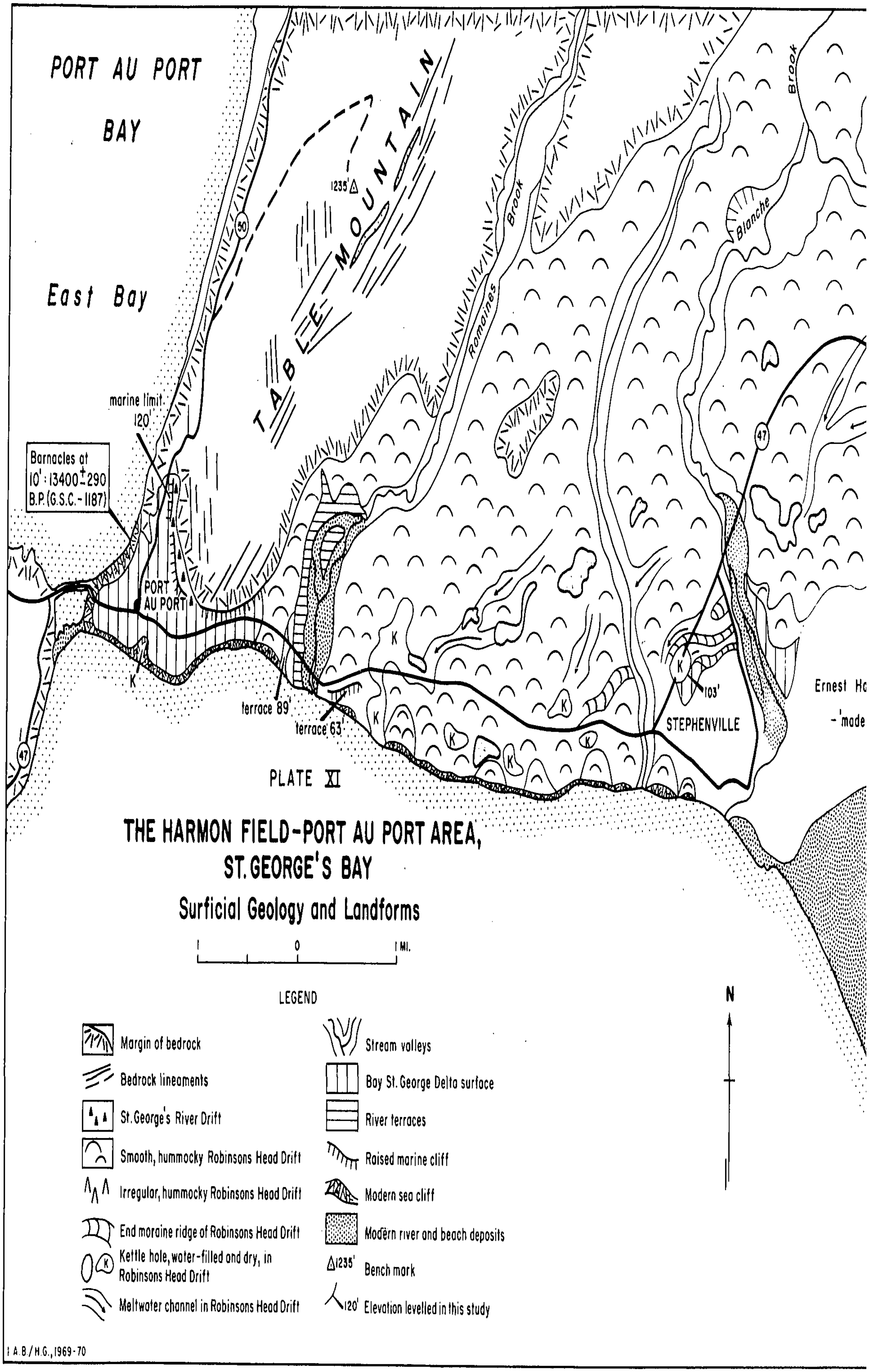
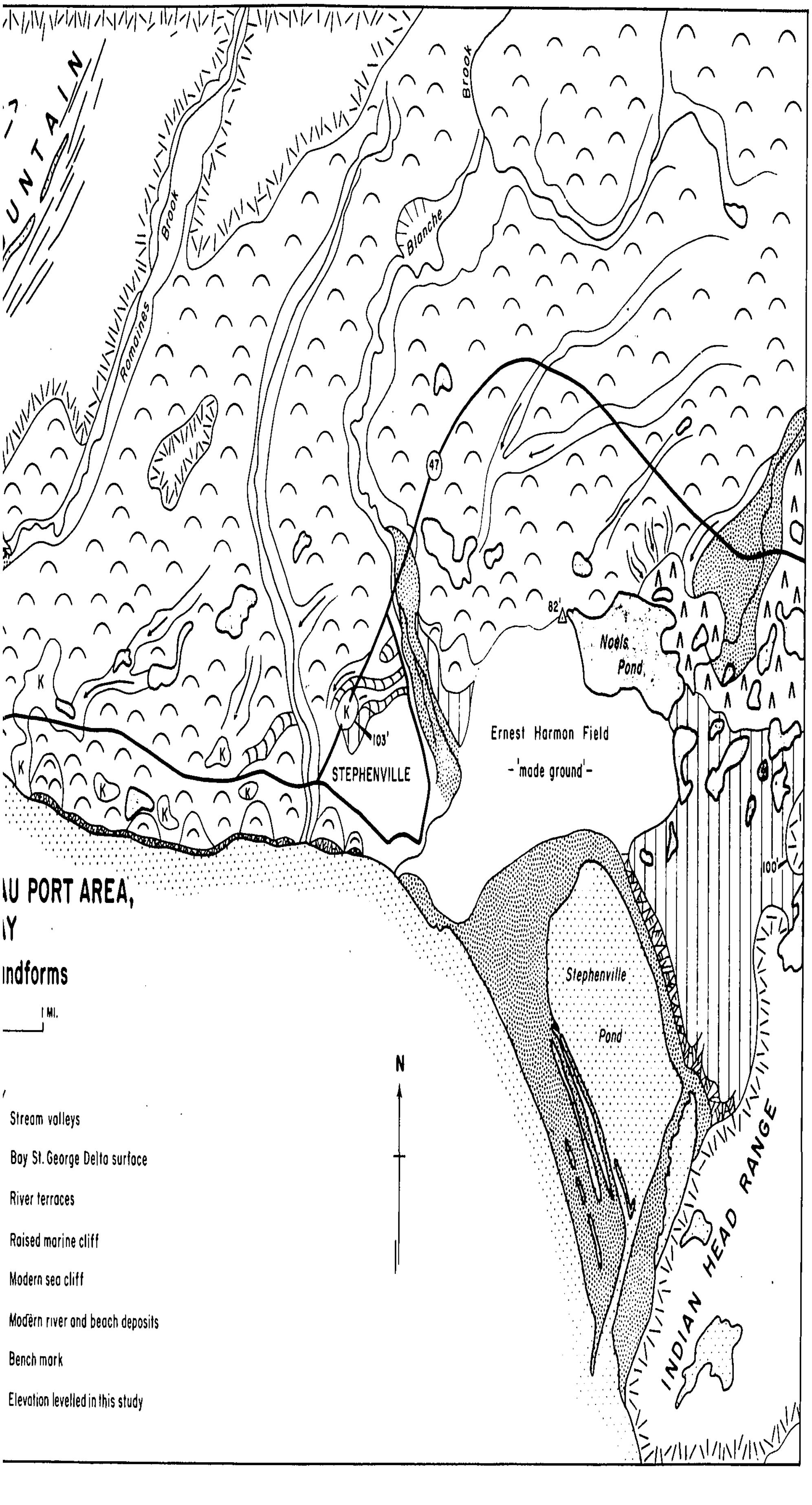


PLATE X
THE HARRYS RIVER LOWLAND
Surficial Geology and Landforms
1
0
1 MI.
LEGEND
Margin of bedrock
Surface outcrop of St. George's River Drift
Smooth, hummocky Robinsons Head Drift
Irregular, hummocky Robinsons Head Drift
Till ridges
Eskers
Sand and gravel "apron" of esker. Outwash plain
Kettle holes, water-filled
Meltwater channel (may be occupied by modern streams)
Modern channel - high-flow boundary
Bay St. George delta surface
"Upper" terrace - west side of lowland
"Second" terrace - west side of lowland
"Lower" terrace - west side of lowland
T Terraces lower than above
Terrace bluff or steep side of modern channel
Floodplain scrolls
Margin of meltwater channel cut in bedrock
Evidence of ice-flow directions on bedrock surface
Δ90 Precise-levelling bench marks, elevation to nearest foot
•130• Points levelled in this study
47 Paved road
I.A.B./G.W.F., 1969-70
N
BLACK
DUCK
Harrys
River
Δ219'
Δ307'
159'
47



PORT AU PORT
BAY
East Bay
TABLE MOUNTAIN
1235'
50
Romaines
Brook
Blanche
Brook
marine limit 120'
Barnacles at
10' : 13400 ± 290
B.P. (G.S.C. - 1187)
PORT AU PORT
K
terrace 89'
terrace 63'
K
103'
STEPHENVILLE
47
47
Ernest Ha
- made
PLATE XI
THE HARMON FIELD-PORT AU PORT AREA,
ST. GEORGE'S BAY
Surficial Geology and Landforms
1 0 1 MI.
LEGEND
Margin of bedrock
Bedrock lineaments
St. George's River Drift
Smooth, hummocky Robinsons Head Drift
Irregular, hummocky Robinsons Head Drift
End moraine ridge of Robinsons Head Drift
Kettle hole, water-filled and dry, in Robinsons Head Drift
Meltwater channel in Robinsons Head Drift
Stream valleys
Bay St. George Delta surface
River terraces
Raised marine cliff
Modern sea cliff
Modern river and beach deposits
1235' Bench mark
120' Elevation levelled in this study
N
I.A.B./H.G., 1969-70



AU PORT AREA,
Y
ndforms

1 MI.

Stream valleys
Bay St. George Delta surface
River terraces
Raised marine cliff
Modern sea cliff
Modern river and beach deposits
Bench mark
Elevation levelled in this study

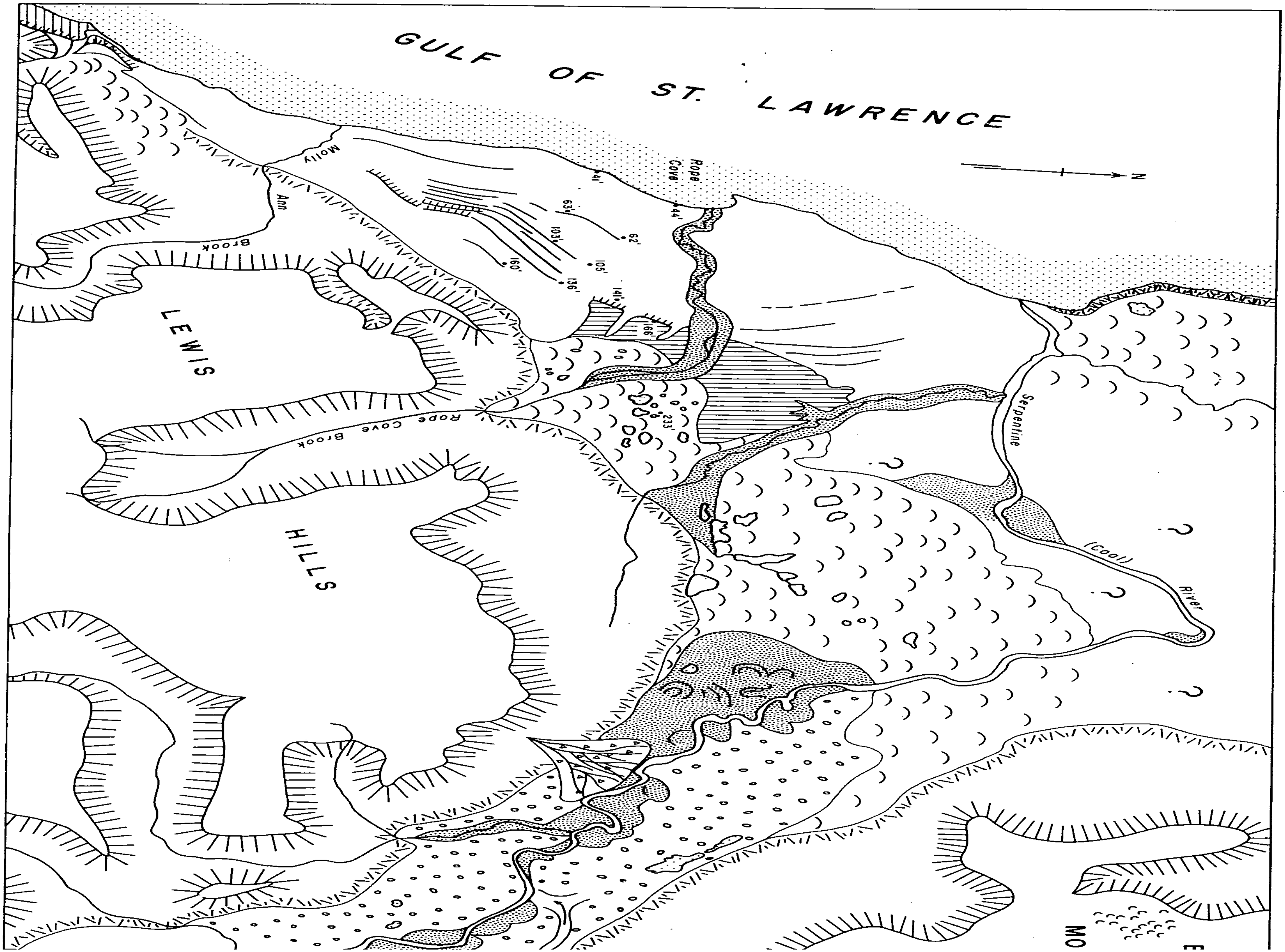

GULF OF ST. LAWRENCE
N
Rope Cove
44'
41'
62'
63'
103'
105'
136'
141'
160'
166'
233'
Molly Ann Brook
Rope Cove Brook
LEWIS HILLS
Serpentine (Coal) River
?
?
?
?
MO
E

PLATE XII

THE LOWER SERPENTINE RIVER SECTION

1 0 1 Mi

LEGEND

- Margin of bedrock
- Free face and talus slope bounding margin of 1750'–2250' plateaus
- Hummocky end moraine
- Kettle holes, water-filled
- Level outwash plain
- Meltwater channel
- Fan deposited by post-glacial Rope Cove Brook
- Alluvial cone
- Wave-cut bluffs
- Marine terrace
- Elevated beach ridges
- Coastal cliffs
- Recent stream channel, with meander traces
- Solifluction lobes on high plateau surfaces
- •141' Elevation levelled in this study
- ? Area unmapped

I.A.B./G.W.F.,1969-70

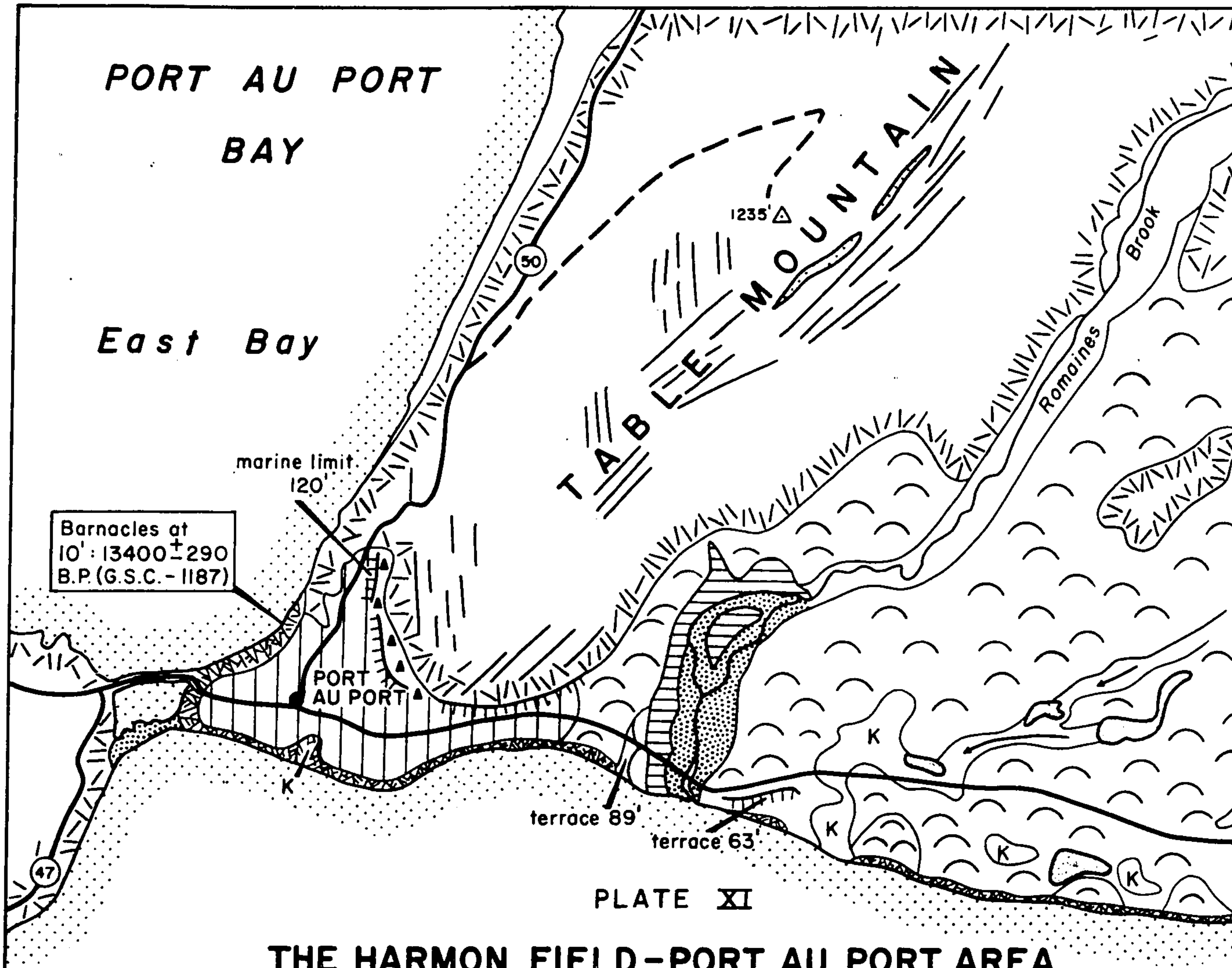


PLATE XI

THE HARMON FIELD-PORT AU PORT AREA, ST. GEORGE'S BAY

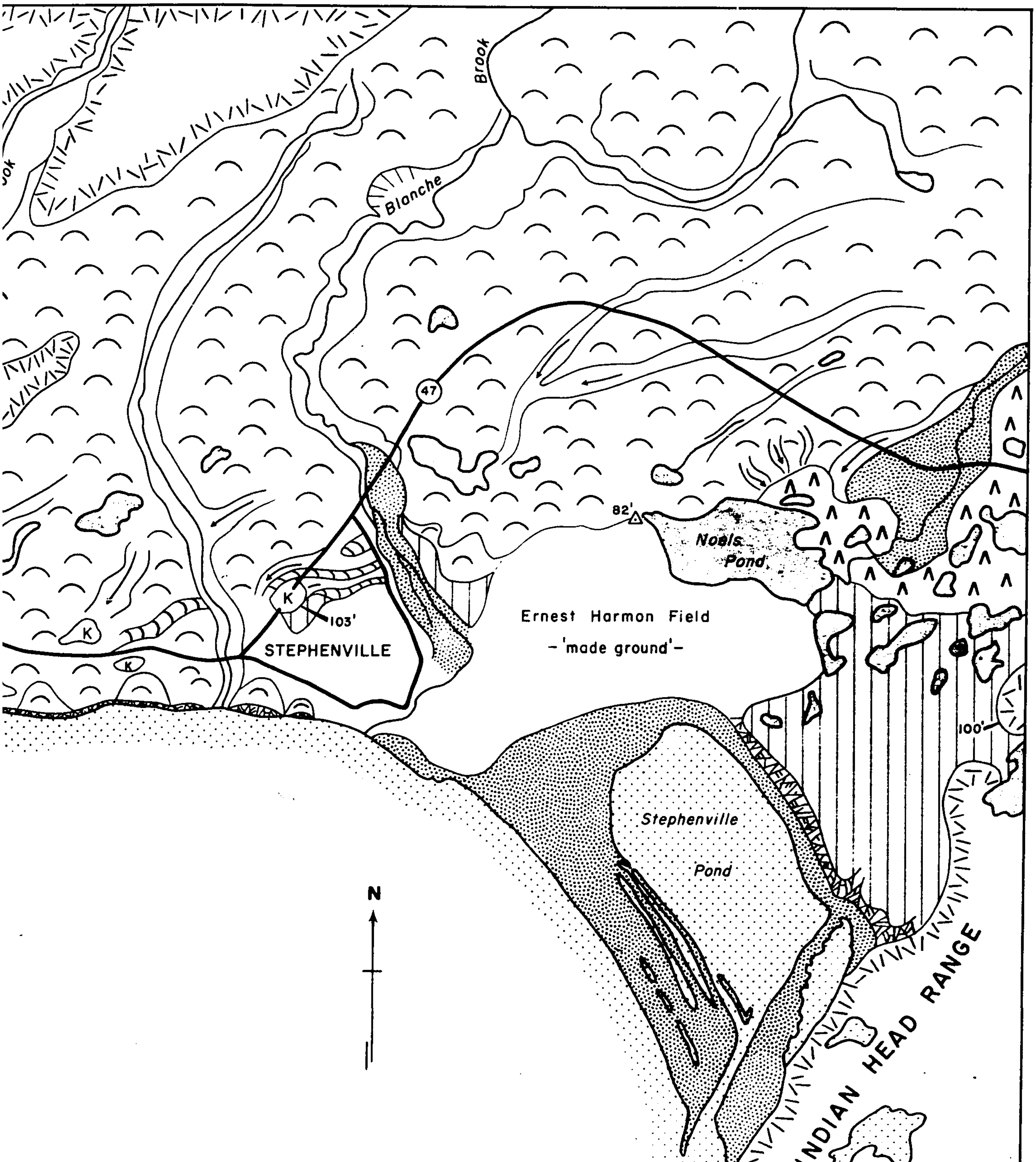
Surficial Geology and Landforms

1 0 1 MI.

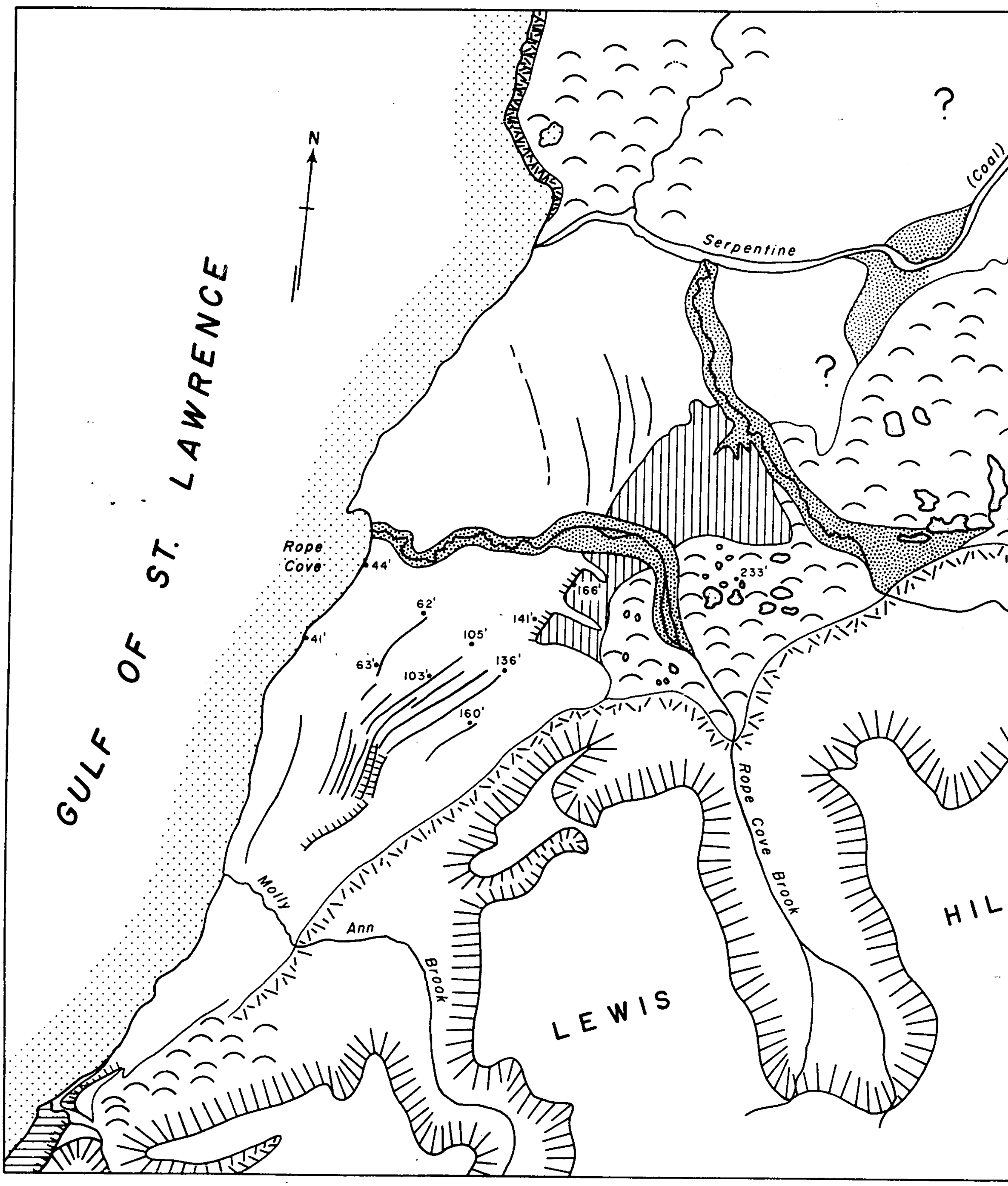
LEGEND

- Margin of bedrock
- Bedrock lineaments
- St. George's River Drift
- Smooth, hummocky Robinsons Head Drift
- Irregular, hummocky Robinsons Head Drift
- End moraine ridge of Robinsons Head Drift
- Kettle hole, water-filled and dry, in Robinsons Head Drift
- Meltwater channel in Robinsons Head Drift
- Stream valleys
- Bay St. George Delta surface
- River terraces
- Raised marine cliff
- Modern sea cliff
- Modern river and beach deposits
- △1235' Bench mark
- 120' Elevation levelled in this study

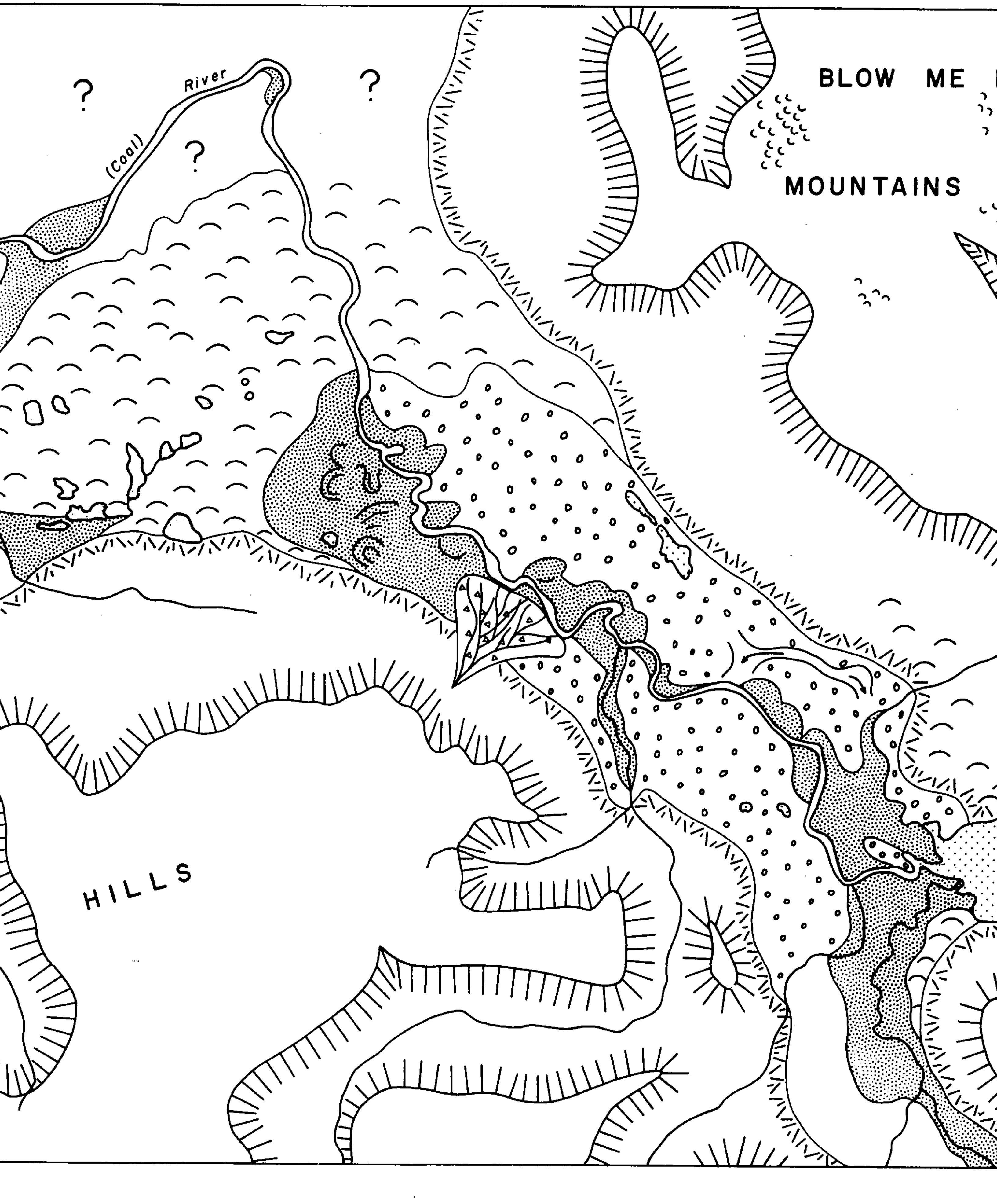
I.A.B./H.G., 1969-70



Brook
Blanche
47
82'
Noels Pond
K
103'
K
K
STEPHENVILLE
Ernest Harmon Field
- 'made ground' -
100'
Stephenville Pond
N
INDIAN HEAD RANGE



GULF OF ST. LAWRENCE
N
Serpentine
(Coal)
?
?
Rope Cove
44'
41'
62'
63'
103'
105'
136'
141'
166'
160'
233'
Rope Cove Brook
Molly
Ann
Brook
LEWIS
HIL



?
River
?
(Coal)
?
BLOW ME
MOUNTAINS
HILLS

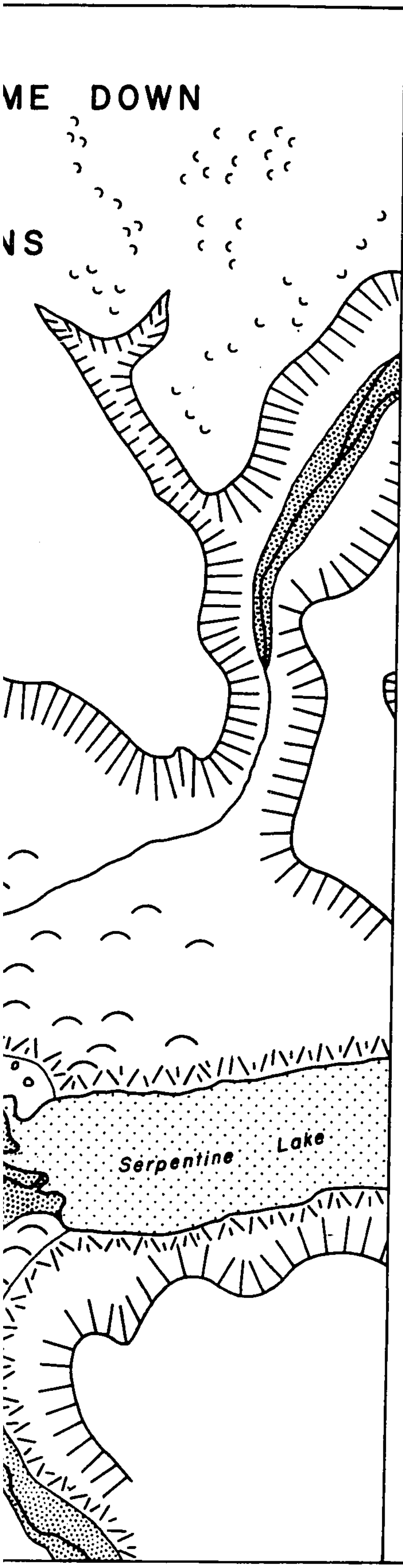


PLATE XII

THE LOWER SERPENTINE RIVER SECTION

1 0 1 Mi.

LEGEND

- Margin of bedrock
- Free face and talus slope bounding margin of 1750'–2250' plateaus
- Hummocky end moraine
- Kettle holes, water-filled
- Level outwash plain
- Meltwater channel
- Fan deposited by post-glacial Rope Cove Brook
- Alluvial cone
- Wave-cut bluffs
- Marine terrace
- Elevated beach ridges
- Coastal cliffs
- Recent stream channel, with meander traces
- Solifluction lobes on high plateau surfaces
- •141' Elevation levelled in this study
- ? Area unmapped

I.A.B./G.W.F., 1969-70